Hurricane Juan (2003). Part I: A Diagnostic and Compositing Life Cycle Study

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ABSTRACT

A detailed analysis of the complex life cycle of Hurricane Juan (in 2003) is undertaken to elucidate the structures and forcings that prevailed over the period leading up to the hurricane’s landfall in Halifax, Nova Scotia, Canada. Despite the presence of easterly wave precursors, Hurricane Juan’s initial development is shown to occur in a baroclinic environment beneath a low-latitude potential vorticity streamer. This feature interacts with a lower-level shear line as the incipient vortex begins to effectively focus ascent and convection. The system undergoes a slow tropical transition over a period of several days as the deep-layer shear over the developing storm decreases. The hurricane is repeatedly perturbed by subsynoptic-scale waves traveling along the leading edge of a large upstream trough. However, Hurricane Juan maintains its tropical structure despite its relatively high formation latitude (28°N) and its northward trajectory. The unusual persistence of the storm’s tropical nature as it propagates northward is of primary interest in this study. In particular, the role of persistent ridging along the east coast of North America is investigated both in high-resolution analyses for Hurricane Juan and in a compositing framework. Dynamic tropopause, quasi-geostrophic, and modified Eady model diagnostics are used to elucidate the interactions between Hurricane Juan and this amplified midlatitude flow. Given the strength and persistence of the anomalous ridge–trough couplet both in the case diagnosis and in the composite fields, the study concludes that the presence of prestorm, high-amplitude ridging along the east coast likely reinforced by diabatic ridging downshear of the storm itself produces an environment both dynamically and thermodynamically conducive to the high-latitude landfall of hurricanes still in the tropical phase.

1. Introduction

Landfalling Atlantic tropical cyclones (TCs) pose a serious threat to low-latitude regions in North America, Central America, and the Caribbean; however, littoral areas north of 40°N are not regularly subject to landfalling hurricanes. Hurricane Juan’s landfall near Halifax, Nova Scotia, Canada (Fig. 1), as a category-2 storm (Saffir–Simpson hurricane scale; Simpson 1974) in September of 2003 claimed the lives of four individuals and represents a relatively rare event worthy of further study. Since 1951, the National Hurricane Center (NHC) best-track archives show that only 17 hurricanes have struck the coastal regions of New England and the Canadian Maritime Provinces, representing less than 5% of the total hurricane activity over the period 1948–2003. No major hurricanes (category 3 or stronger) have made landfall along the eastern seaboard north of 40°N over this period, although both hurricanes Ginny (in 1963) and Gerda (in 1969) struck the coast with estimated near-surface winds of 48 m s⁻¹. The average intensity of the hurricanes that make landfall north of 40°N is 37 m s⁻¹ (category 1). These statistics, however,
do not imply that the effects of hurricane landfall at higher latitudes along the east coast of North America are any less severe than those in other areas.

With a 3-yr return period as diagnosed from the best-track records, it is precisely the relative infrequency of hurricane landfall in New England and the Maritimes that makes these storms a serious threat to these regions. Following the passage of the intense Hurricane Gloria (in 1985) through Canadian waters, the Canadian Hurricane Centre (CHC) was established to facilitate the forecasting of TCs reaching higher latitudes in the western Atlantic Ocean. Both the CHC and the NHC forecast for hurricane tracks and intensities in their respective regions of responsibility; however, to date the CHC issues significant weather warnings (wind, precipitation, wave and storm surge) rather than hurricane warnings similar to those produced by the NHC. These more northerly coastal areas also lack the experienced, extensive hurricane emergency management, evacuation, and mitigation infrastructures necessary in regions more commonly affected by major landfalling hurricanes. For these reasons, the risk of hurricane landfall to higher-latitude coastal areas along the eastern seaboard is particularly great and the study of such an event may yield valuable insights into the environment, structures, and motion of a potentially high-impact hurricane.

To develop an enhanced understanding of Hurricane Juan, a detailed diagnosis of the storm’s complex life cycle is required. In this context, a complex life cycle refers to one that involves significant tropical and extratropical stages. The forcings that drive the formation of tropical depressions from easterly wave precursors have been investigated extensively by Erickson (1963), Carlson (1969a,b), Burpee (1972), Shapiro (1977), Tuleya and Kurihara (1981), Kurihara and Kawase (1985), Thorncroft and Hodges (2001), Karyampudi and Pierce (2002), and many others. However, the role played by the extratropics in such developments has received attention only recently (Bosart and Bartlo 1991; Montgomery and Shapiro 1993; Elsner et al. 1996; Bracken and Bosart 2000; Davis and Bosart 2001, 2003, 2004).
Davis and Bosart (2001, 2002) show that midlatitude forcings may initially influence tropical developments by focusing ascent in regions consistent with quasigeostrophic (QG) theory. Thereafter, mesoscale and convective processes dominate as the incipient TC progresses through the remaining steps of the three-stage tropical transition (TT) process. In more extreme cases, dubbed strong extratropical cyclone transitions by Davis and Bosart (2004), initially baroclinic systems undergoing TT can evolve directly into tropical storms (Bosart and Bartlo 1991). Such transitions have been shown to take place in environments whose thermal structures are conducive to traditional tropical genesis (Palmén 1958; Gray 1968; Rotunno and Emanuel 1987), albeit in regions where vertical wind shears may exceed critical values suggested by DeMaria et al. (2001). Davis and Bosart (2003, 2004) show that a diabatically driven occlusive process characterized by strong upper-level outflow and vertical redistribution of potential vorticity (PV) is responsible for reducing shear values to below the tropical genesis threshold.

The importance of midlatitude structures to TC motion and intensity changes has been documented by Holland and Merrill (1984), Molinari and Vollaro (1990), Molinari et al. (1995, 1998), Hanley et al. (2001), Kimball and Evans (2002), and others. Owing to the relatively high formation latitude, the predominantly northward trajectory of Hurricane Juan, and the large latitudinal extent of the midlatitude circulation over the period of interest, these extratropical influences were strong throughout the TC life cycle. The extratropical transition (ET) of Atlantic hurricanes is a relatively common occurrence (DiMego and Bosart 1982; Hart and Evans 2001; Jones et al. 2003), with between two and three transitioning storms influencing the Canadian Maritimes each year (Joe et al. 1995). On average, 50% of Atlantic TCs of hurricane strength undergo transformations to extratropical systems. The tropical–midlatitude interactions that occur during ET have been investigated extensively as described in the review paper by Jones et al. (2003).

In this study, an attempt has been made to consolidate some of the theories and conceptual models developed through previous investigations of the genesis, TT, TC intensity change, and ET problems. The case-based framework of this research permits a detailed investigation of the processes and structures present during Hurricane Juan’s complex life cycle. The study begins with a description of the datasets and analysis methods in section 2. The results of a compositing study involving landfalling hurricanes in New England are presented in section 3. In section 4 an analysis of Hurricane Juan’s life cycle from its easterly wave precursor to its landfall in the Canadian Maritimes is presented. The study concludes with a discussion of the findings in section 5.

2. Data and methods

This study makes use of gridded and tracking data from a variety of sources. Tropical cyclone track and intensity estimates are obtained from the NHC best-track archives, but analysis of these tracks is restricted to the 1948–2003 period consistent with the extent of the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (NNR) dataset (Kalnay et al. 1996). The 2.5° gridded NNR is used in the compositing portion of the study presented in section 3. For consistency, and to ensure independence from the case itself, all climatologies are based on August–October analyses over the 55-yr period from 1948 to 2002.

The diagnosis of Hurricane Juan’s development from easterly wave precursors (section 4a) is based on full-resolution (1° gridded) 6-hourly analyses from NCEP’s Global Data Assimilation System (GDAS; more information available online at http://www.emc.ncep.noaa.gov/gmb/gdas/; Caplan et al. 1997). Use of this dataset for the genesis diagnostics is necessary because of the large spatial extent of the features of interest, especially those in the eastern tropical Atlantic. All other primary diagnostics (section 4) are performed using 12-hourly data from NCEP’s 12-km griddedEta Data Assimilation System (EDAS; Black 1994; Rogers et al. 1996) on the NCEP 221 grid. This 32-km Lambert conformal output grid has the advantage of encompassing virtually the full domain of the EDAS without unacceptable degradation of the analysis resolution.

The EDAS analyses are used as the primary dataset for the diagnostic component of this study despite intermittent difficulties with the system’s ability to accurately depict tropical cyclones. In particular, experience suggests that the EDAS has trouble representing intense systems with short length scales despite its high resolution. The effects of this shortcoming are minimal on this study for two reasons. First, although a compact storm throughout its life cycle, Hurricane Juan never exceeded category-2 intensity (Simpson 1974). As will be shown in section 4d, the EDAS analyses prove capable, in this case, of representing the storm’s strong lower-level vorticity maximum. Second, the goals of this study focus on the environmental forcings and interactions that occur during the storm’s lifetime rather than on the details of the analyzed hurricane’s structure.
and intensity. The high resolution of the EDAS analysis permits valuable discussion regarding robust, dynamically significant features in the environmental flow that are not well resolved by coarser global analyses. Indeed, a set of secondary diagnostics—only a subset of which are shown—shadows the EDAS analyses using the 1° GDAS grids described above. These diagnoses demonstrate that the features and processes described in section 4 are not simply artifacts of EDAS analysis irregularities.

Anomaly fields are computed relative to the 55-yr August–October climatology mentioned above, unless otherwise noted. The primary exception to this methodology involves the Eady model–type diagnostics (Eady 1949), for which perturbations are taken relative to a basic state consistent with the constraints of Eady’s simplified model. A detailed description of the idealized background is included in section 4c.

3. Climatology of high-latitude landfalling hurricanes

Hurricane Juan’s acceleration on a northward trajectory as it neared the Nova Scotia coast (Fig. 1) suggests a strengthening southerly steering flow along the eastern seaboard of the United States. This is indicative of the strong trough–ridge couplet centered over the area that is described in detail in section 4. However, the extent to which this flow pattern represents a necessary feature germane to high-latitude landfalling hurricanes remains unclear. To address this issue in a general manner, a basic compositing study is presented in this section. The goals of this study are twofold: to diagnose the presence of east coast ridging over multiple events; and to determine to what extent this ridging is a result of the diabatically driven hurricane outflow, or a preexisting feature of the midlatitude environment.

The storms included in the composite had to fulfill the following requirements:

1) occur between 1948 and 2002 (inclusive);
2) make landfall north of 40°N as hurricanes (according to NHC best track and verified by the analysis); and
3) have a motion vector between northwestward and northeastward.

Criterion 1 restricts the composite storms to the period covered by the NNR while removing Juan from the study to ensure the independence of the mean state. Both the structures and the intensities of the TCs are addressed by criterion 2. Criterion 3 ensures that the storms are on a northward trajectory, implying that re-curvature is not complete and that each storm is likely to penetrate significantly into higher latitudes. Only eight storms (listed in Table 1) meet all three criteria and are included in the composite dataset. The T − 00h compositing time is taken to be the last 6-hourly NHC best-track report prior to landfall. All grids are interpolated to a common reference point of 43°N, 68°W (in the Gulf of Maine) before compositing, making the geographical references in this section somewhat arbitrary; however, no displacements of greater than 5° of latitude were required in either direction for any of the composite members. A set of thumbnail images showing the uninterpolated 500-hPa height fields for Juan, the composite mean, and each of the eight cases (as labeled) is presented in Fig. 2. The presence of strong ridging over western North America and along the eastern seaboard (and the trough implicit between them) at T − 00h confirms the instantaneous stability of the composite since the members contain similar structures at the compositing time. The persistence of these features throughout the 48 h leading up to T − 00h (Fig. 2) further suggests that there are extensive dynamical commonalities in the development of the midlatitude flow among the members.

The amplitude and significance of the midlevel east coast ridging at T − 00h is clearly demonstrated in Fig. 3e. Height anomalies at 500 hPa in excess of 15 dam are present north of Newfoundland and a broad region of statistically significant anomalous ridging extends from the Maritimes to the southern tip of Greenland. To the west of the hurricane (denoted by a hurricane symbol in Fig. 3e) troughing dominates over a large area from the coast to the Great Lakes. The first objective of this compositing study has thus been achieved; east coast ridging is a dominant signal over

Table 1. Hurricanes used for the compositing study presented in section 3. Last prelandfalling reports (T − 00h), intensities, and trajectories are determined from NHC best-track data. Subjectively defined trajectory abbreviations are (N) northward, (NNE) north-northeastward, and (NE) northeastward.

<table>
<thead>
<tr>
<th>Year</th>
<th>Name</th>
<th>T − 00h</th>
<th>Intensity (m s⁻¹)</th>
<th>Trajectory</th>
</tr>
</thead>
<tbody>
<tr>
<td>1953</td>
<td>Carol</td>
<td>1200 UTC</td>
<td>43</td>
<td>NNE</td>
</tr>
<tr>
<td>1954</td>
<td>Carol</td>
<td>1800 UTC</td>
<td>43</td>
<td>NNE</td>
</tr>
<tr>
<td>1955</td>
<td>Edna</td>
<td>1800 UTC</td>
<td>43</td>
<td>NNE</td>
</tr>
<tr>
<td>1960</td>
<td>Donna</td>
<td>1800 UTC</td>
<td>43</td>
<td>NNE</td>
</tr>
<tr>
<td>1963</td>
<td>Ginny</td>
<td>1200 UTC</td>
<td>43</td>
<td>NNE</td>
</tr>
<tr>
<td>1969</td>
<td>Gerda</td>
<td>0000 UTC</td>
<td>43</td>
<td>NNE</td>
</tr>
<tr>
<td>1975</td>
<td>Blanche</td>
<td>1200 UTC</td>
<td>43</td>
<td>NNE</td>
</tr>
<tr>
<td>1990</td>
<td>Bertha</td>
<td>0000 UTC</td>
<td>43</td>
<td>NNE</td>
</tr>
</tbody>
</table>
Fig. 2. Analyzed 500-hPa geopotential heights (solid contours at 6-dam intervals) and departures from the composite mean (light dashed contours at 3-dam intervals with no zero contour) for (i) Juan, (ii) the composite mean, and (iii–x) the composite members as labeled. Fields are shown at (a) T = 00h, (b) T = 24h, and (c) T = 48h.
multiple events as anticipated based on the presence of anomalous southerly steering flow (implicit in criterion 3 above).

To address the second goal, the dynamics associated with the east coast ridging must be evaluated in the compositing framework. Two days before the hurricane’s landfall (T = 48h, Fig. 3a), ridging over the Rocky Mountains disrupts a primarily zonal flow over the continent. Weak anomalous ridging is also beginning to appear along the eastern seaboard (Fig. 3b) but is not yet a statistically significant feature. Twenty-four hours later (T = 24h, Fig. 3c), the ridge over western North America reaches its maximum amplitude with 500-hPa heights up to 12 dam above their climatological values. Ridge building is also evident over the Maritimes in advance of the tropical system whose composite representation appears southeast of Cape Hatteras. As the western ridge decays, the downstream trough over the Great Lakes and the east coast ridge intensify (Fig. 3d). This downstream propagation suggests that Rossby wave dynamics may play an important role in organizing the large-scale flow. By T = 00h (Fig. 3e), this trough–ridge couplet is positioned to produce the anomalous southeasterly steering flow in the hurri-

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**Fig. 3.** Composite 500-hPa heights (solid lines at 6-dam intervals) and anomalies from the 55-yr mean (dotted lines at 3-dam intervals above +6 dam and below −6 dam). Shading represents statistical significance, with light and dark shading for the 90% and 95% significance levels, respectively. The times in all panels represent departures from the last NHC best-track time before landfall (T = 00h): (a) T = 48h, (b) T = 36h, (c) T = 24h, (d) T = 12h, and (e) T = 00h.
cane’s vicinity, responsible for the northward propagation of the tropical system.

The anomalous amplitude of the east coast ridge–trough couplet at T = 00h (Fig. 3e) is undoubtedly increased by the diabatic outflow from the composite hurricane; however, the coherence of the synoptic-scale flow over the 48-h period prior to landfall suggests that it has midlatitude origins. Furthermore, the Rossby wave–like propagation of the western ridge at T = 48h suggests that large-scale forcings contribute to the rearrangement of the synoptic-scale flow. The second goal of this compositing study is thus left only partially addressed since the role of the hurricane’s outflow in further developing the ridge cannot be neglected based on this analysis. A more detailed diagnosis of the composite structures and dynamics, particularly as they pertain to Rossby wave train propagation and energetics, is beyond the scope of the current study and will be the focus of a subsequent investigation.

4. Hurricane Juan

This section comprises a set of separate analyses of the varied components of Hurricane Juan’s complex life cycle. In section 4a, the easterly wave (EW) precursor that plays a critical role in the initiation of the hurricane is tracked across the equatorial Atlantic. Section 4b constitutes a descriptive analysis of Juan’s TT, deepening, and landfalling stages from a dynamic tropopause perspective. Interpretations of the developmental and QG forcings are reserved for sections 4c and 4d, respectively. Through its implicit separation of tropical and extratropical dynamics, the Eady model (Eady 1949) perspective employed in section 4c serves to elucidate the TT of Hurricane Juan, a result of the favorable combination of a strong baroclinic precursor and the EW described in section 4a. Section 4d focuses on the importance of synoptic-scale ascent forcing during the development and deepening stages of Juan’s life cycle through an analysis of these steps from a QG perspective. The figures and text in sections 4b and 4c make use of a number of symbols for tracking important features involved in Hurricane Juan’s evolution. A reference list for these symbols can be found in Table 2.

a. Precursors

The genesis of Hurricane Juan can be traced back to a strong EW crossing over the Liberian coast on 11 September 2003 (Fig. 4a). After moving offshore, the EW rapidly intensified near 6°N, 14°W as indicated by the increase in the 850-hPa relative vorticity by 1200 UTC 11 September. Over the next 36 h, this EW interacts and eventually merges with another strong EW moving off of the Liberian coast at 0000 UTC 12 September (not shown). By 0000 UTC 14 September, a single coherent wave is readily identifiable in the 850-hPa vorticity field located near 8°N, 30°W with a maximum relative vorticity value of approximately 6 × 10⁻⁶ s⁻¹ (Fig. 4b). The EW travels west-northwestward over the next 6 days, arriving near 14°N, 52°W by 0000 UTC 20 September (Fig. 4c). At this point, the system’s structure becomes more complex as it starts to interact with an upper-tropospheric trough that has penetrated into the Caribbean Sea (to be discussed in detail in section 4b). The reflection of this trough in the vorticity field is the extension in the axis of cyclonic vorticity toward the Lesser Antilles (not shown). The EW then accelerates northward in response to its interaction with the upper-tropospheric trough, ending up near 20°N, 53°W by 0000 UTC 21 September (Fig. 4c). The evolution of the tropical wave and its impact on the eventual genesis of Juan becomes more convoluted after this point. As will be discussed in more detail in the following sections, it is unclear whether the EW itself develops into Juan or whether the EW acts mainly to enhance the convection in the vicinity of the upper-level trough.

b. The dynamic tropopause and tropical transition

Figure 5 shows θ_T, the potential temperature θ on the dynamic tropopause (DT) as defined by the 2 PVU (where 1 PVU = 10⁻⁶ m² K kg⁻¹ s⁻¹) surface. The utility of this representation is that θ is a conserved quantity on material PV surfaces in the absence of frictional and diabatic effects. Therefore, areas in which

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Section(s)</th>
<th>Figure(s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Upper-tropospheric trough</td>
<td>4b</td>
<td>5</td>
</tr>
<tr>
<td>2</td>
<td>Secondary shortwave upper-tropospheric trough</td>
<td>4b</td>
<td>5</td>
</tr>
<tr>
<td>3</td>
<td>Upper-tropospheric trough</td>
<td>4b</td>
<td>5</td>
</tr>
<tr>
<td>4</td>
<td>Upstream upper-tropospheric trough</td>
<td>4b</td>
<td>5</td>
</tr>
<tr>
<td>A</td>
<td>Perturbation on equatorward arm of jet stream</td>
<td>4b, 4c</td>
<td>5, 10</td>
</tr>
<tr>
<td>J</td>
<td>Hurricane Juan</td>
<td>4b, 4c</td>
<td>5, 9, 10</td>
</tr>
<tr>
<td>L</td>
<td>Incipient cyclone (low pressure center)</td>
<td>4c</td>
<td>9, 10</td>
</tr>
<tr>
<td>O</td>
<td>Convective outflow from Hurricane Juan</td>
<td>4b, 5</td>
<td></td>
</tr>
<tr>
<td>T</td>
<td>PV tail (cutoff) from trough 1</td>
<td>4b, 4c</td>
<td>5, 9, 10</td>
</tr>
<tr>
<td>W</td>
<td>Warm dynamic tropopause anomaly (ridge)</td>
<td>4b</td>
<td>5</td>
</tr>
</tbody>
</table>
local free-atmospheric changes in $\theta_{DT}$ cannot be attributed to advection are likely experiencing redistribution of PV through convection. For a more complete discussion of the utility and interpretation of DT maps, the reader is referred to Morgan and Nielsen-Gammon (1998).

At 0000 UTC 21 September, a pronounced minimum of $\theta_{DT}$ (cyclonic PV) is centered near 43$^\circ$N, 46$^\circ$W (labeled “1” in Fig. 5a). A region of low $\theta_{DT}$ extends to the southwest of this feature as a PV “tail” or region of cyclonic shear in the DT wind field. A closed circulation at the end of this tail is located near 20$^\circ$N, 70$^\circ$W (labeled “T” in Fig. 5). It is this feature with which the EW discussed above interacts and eventually induces the genesis of Juan. Another trough (labeled “2” in Fig. 5) is moving off of the eastern seaboard of the United States. Initially, the two troughs (1 and 2) are separated by an elevated DT with values of 365–370 K throughout the Maritimes. Overall, the pattern can be described as highly amplified with strong meridional flow. Over the next 24 h (0000 UTC 21 September–0000 UTC 22 September), the region of high-$\theta_{DT}$ air dissipates as the troughs begin to merge (Figs. 5b,c), resulting in increased northerly flow on the back side of the combined trough with DT wind speeds increasing to 35 m s$^{-1}$. Another amplified trough (labeled “3” in Fig. 5), with a well-defined lower-level front (not shown), is progressing eastward across the Great Lakes at 0000 UTC 23 September (Fig. 5e). A large region of convection associated with this baroclinic zone helps to produce a diabatically enhanced ridge over the eastern seaboard in Figs. 5d–f.

By 0000 UTC 24 September (Fig. 5g), the ridging associated with the frontal convection has begun to fold over the low-latitude PV anomaly (T) now located just east of the Bahamas. An elongated lobe of low-level relative vorticity persists to the east beneath the upper-level anomaly and is collocated with a series of convectively active regions evident in water vapor (WV) imagery (Fig. 6a). Between 0000 UTC 24 September and 0000 UTC 25 September (Figs. 5g–i) a southward extension of the cold trough over North America (labeled “A” in Fig. 5) approaches the closed cyclonic circulation (T), now located near 25$^\circ$N, 65$^\circ$W (Fig. 5i). Meanwhile, the zonally oriented strip of low-level vorticity below the anomaly begins to buckle, increasing its maximum vorticity to values exceeding $24 \times 10^{-5}$ s$^{-1}$ in the northeast quadrant of the DT perturbation (T).

Note that this is a region of cold $\theta_{DT}$ advection (cyclonic advection of PV), suggesting that the increase in low-level vorticity is at least partially a result of baroclinic processes. In turn, this forcing organizes convection in the area and initiates a positive feedback cycle by further enhancing vorticity through column stretching. Water vapor imagery for 0000 UTC 25 September (Fig. 6b) shows a rapid local increase in convective activity and the NHC declares Juan a tropical storm at this time.

The development of Juan proceeds rapidly over the next 24–36 h. By 1200 UTC 25 September (Fig. 5j), the previously closed DT circulation (T) has opened and moved northward to a position near 29$^\circ$N, 61$^\circ$W in response to the shortwave trough (A) approaching it from the west. The proximity of the two features results in the development of a negatively tilted trough axis. This orientation shortens the wavelength between the

![Fig. 4. Mosaic of 850-hPa relative vorticity in Juan's precursor easterly wave (contoured and shaded at $8 \times 10^{-6}$ s$^{-1}$ intervals above zero only) at (a) 0000 UTC 11 Sep, (b) 0000 UTC 14 Sep, (c) 1200 UTC 17 Sep, (d) 0000 UTC 20 Sep, and (e) 0000 UTC 21 Sep 2003.](image-url)
trough and the ridge, and implies an increase in the magnitude of differential cyclonic vorticity advection and the associated QG forcing for ascent. The EDAS analysis at 1200 UTC 25 September (Fig. 5j) identifies several local maxima in the lower-level vorticity field, and although the NHC best-track position suggests that the vorticity center near 31°N, 62°W is the incipient Juan, precise identification of the seedling vortex is difficult at this time. However, by 0000 UTC 26 September (Fig. 5k), a well-formed circulation center with maximum absolute vorticity exceeding 48 × 10⁻⁵ s⁻¹ (labeled “J” in Fig. 5) is located on the eastern side of the midlatitude trough (A). At the same time, θDT values in the ridge ahead of Juan have increased 5–10 K in response to the developing convective outflow of Juan (labeled “O” in Figs. 5k–m). The location of Juan’s low-level circulation center near the maximum θDT gradient, and under a region of cyclonic PV advection, suggests that Juan is still responding in part to baroclinic forcing, even though the NHC has declared Juan a hurricane by this time. Moreover, the convective outflow from Juan contributes to the QG ascent forcing by amplifying the upstream ridge. Water vapor imagery for 0000 UTC 26 September (Fig. 6c) supports the contention that Juan is still, at the very least, a hybrid storm since a comma-head/dry slot structure is evident and similar in nature to that shown by Bosart and Bartlo (1991) during the TT of Hurricane Diana (in 1984).

By 1200 UTC 26 September (Fig. 5l), the upper- and lower-level circulation centers of Juan (T and J, respectively) have become vertically aligned (Fig. 5h). The highest θDT air (O) is still located north and east (downstream) of Juan, suggesting at least moderate shear values (between 5 and 10 m s⁻¹ in the 200–850-hPa layer) near the storm center. The upstream midlatitude trough (A) continues to approach Juan, causing the hurricane to move northward. The proximity of this cold-core trough to Juan’s diabatically enhanced ridge produces an enhanced θDT gradient to the west of the storm that strengthens the jet in this region by 15 m s⁻¹ in 12 h to over 25 m s⁻¹ by 1200 UTC 26 September. This strengthened flow provides an enhanced outflow channel for mass evacuation northwest of the hurricane and the low-level circulation center of Juan responds by strengthening, with maximum relative vorticity values exceeding 60 × 10⁻⁵ s⁻¹ by 0000 UTC 27 September (Fig. 5m). The extent to which upper-level deformational processes driven by the interaction of the environmental flow and the hurricane’s circulation contribute to the intensification of the vortex remains a subject for future study. Water vapor imagery at this time (Fig. 6d) shows a tropical structure with a well-defined eye and cold, dense central overcast. Another important feature in the 0000 UTC 27 September analysis is the high (>375 K) θDT air (labeled “W” in Figs. 5m–r) preceding the next highly amplified trough over the western Great Lakes (labeled “4”). This feature ultimately establishes an anomalously warm (“tropical”) high-latitude environment for Juan as the storm moves northward.

At 1200 UTC 27 September (Fig. 5n), a band of low-level vorticity develops between Juan (J) and the center of the nearby upper-level trough (A) and Juan makes an abrupt cyclonic track change (Fig. 1). Water vapor imagery for 0000 UTC 28 September (Fig. 7a) continues to show a well-defined eye, with dry air encroaching upon Juan from the west and south in association with the weakening trough axis. By 1200 UTC 28 September, Hurricane Juan has resumed its northward track and has moved into a region of elevated θDT (>360 K as shown in Fig. 5p). The unusual aspect of this evolution is that the tropical, high-θDT environment in which Juan is embedded for the remainder of the tropical phase of its life cycle is not native to the storm itself. Instead, this environment is a direct result of the highly amplified synoptic-scale pattern. This flow appears to explain why Juan was able to maintain its strength and tropical characteristics at such high latitudes. Between 1200 UTC 28 September and 1200 UTC 29 September (Figs. 5p–r) Juan accelerates northward in the large-scale southerly flow preceding the next trough (4), increasing the impact of the hurricane’s strong winds and storm surge in the eastern quadrant of the fast-moving cyclone, while at the same time minimizing its residence time over cooler waters.

As the hurricane makes landfall near Halifax at 0300 UTC 29 September (Fig. 1), it becomes engulfed in the large θDT ridge (W) now covering much of the Maritimes (Figs. 5q–r). The overwhelming scale of the approaching midlatitude trough (4) relative to that of Juan essentially precludes any significant intensification during ET. The large areal extent of the high-shear region associated with the trough PV anomaly subjects the approaching storm to an extended period of unfavorable forcing (Molinari et al. 1998; Hanley et al. 2001) that results in the destruction of the vortex shortly after 1200 UTC 29 September over Labrador, Canada (Fig. 1).

A parallel set of dynamic tropopause diagnostics has been constructed from the 1.0° GDAS analyses as described in section 2. This methodology permits independent verification of the features described throughout this section by direct comparison of Figs. 5 and 8. The daily 0000 UTC analyses from the coarser-resolution GDAS from 0000 UTC 24 September to
Fig. 5. The DT EDAS analysis of $\theta_{D}$T (colors as indicated on the attached color bar at 5-K intervals) and winds (barbs with pennants and barbs corresponding to 25 and 5 m s$^{-1}$ increments, respectively). Relative vorticity at 850 hPa is plotted at $6 \times 10^{-5}$ s$^{-1}$ intervals above $12 \times 10^{-5}$ s$^{-1}$ in solid black lines. The DT is here taken to be the 2-PVU surface. Panels are plotted sequentially at 12-h intervals from 0000 UTC 21 Sep 2003 and correspond to (a) 0000 UTC 21 Sep, (b) 1200 UTC 21 Sep, (c) 0000 UTC 22 Sep, (d) 1200 UTC 22 Sep, (e) 0000 UTC 23 Sep, (f) 1200 UTC 23 Sep, (g) 0000 UTC 24 Sep, (h) 1200 UTC 24 Sep, (i) 0000 UTC 25 Sep, (j) 1200 UTC 25 Sep, (k) 0000 UTC 26 Sep, (l) 1200 UTC 26 Sep, (m) 0000 UTC 27 Sep, (n) 1200 UTC 27 Sep, (o) 0000 UTC 28 Sep, (p) 1200 UTC 28 Sep, (q) 0000 UTC 29 Sep, and (r) 1200 UTC 29 Sep 2003. Details concerning specific annotations are contained in the text.
0000 UTC 29 September (Fig. 8) show that the evolutions of the primary structures (labeled consistently with the EDAS analyses shown in Fig. 5) are highly similar to those described earlier in this section for the EDAS. One exception to this observation lies in the strength and structure of the diabatically enhanced lower-level vorticity streamer important for Juan’s development (cf. the vorticity contours in Figs. 5i and 8b). This discrepancy between the EDAS and the GDAS analyses is discussed in detail in section 4d. Overall however, the consistency between the analyses provides confidence that the diagnoses in this section and
those that follow focus on real atmospheric structures and not on artifacts of the EDAS dataset.

c. Coupling potential and the Eady model

In this section, the genesis of Juan will be further investigated using diagnostics that are descriptive of baroclinic development. As such, the discussion will be limited to the storm’s development and intensification phases, covering the period from 0000 UTC 24 September to 1200 UTC 26 September. Figure 9 displays the pressure on the DT, the low-level wind field, and the coupling potential, defined as the difference be-
between $\theta_{DT}$ and the equivalent potential temperature $\theta_e$ at 925 hPa (Bosart and Lackmann 1995). Low, or negative, values of the coupling potential indicate regions of reduced bulk column stability. In such an environment, vertical communication between PV anomalies is more effective because of the increased Rossby penetration depth. One might therefore expect these areas to be more amenable to cyclogenesis.

At 0000 UTC 24 September (Fig. 9a), an inverted shear line (denoted with dashes in the figure) is evident in the low-level wind field stretching east-northeast from a weak cyclonic vortex located near 25°N, 69°W (identifiable in the lower-level wind field and labeled as feature “L” in Fig. 9). This feature is collocated with both an axis of low coupling potential and a local maximum in $P_{DT}$ (labeled feature “T” in Fig. 9). The minimum in coupling potential here is achieved both through a $\theta_{DT}$ minimum (see Fig. 5g) and a maximum in the $\theta_e$ at lower levels as expected over the warm sea surface (not shown). Recall that the water vapor imagery (Fig. 6a) indicates a series of convective bursts aligned along this low-level trough. The trough maintains its linear structure over the next 12 h as the upper-level PV anomaly to its southwest (T) becomes better defined with a maximum in $P_{DT}$ of 300 hPa east of the Bahamas by 1200 UTC 24 September (Fig. 9b). Even though the NHC has already identified a tropical depression at this time, the only closed circulation evident in the EDAS analysis wind field is located well to the southwest of the NHC best-track center. This discrepancy in the placement of a weak

1 The influence of a PV anomaly decays with a vertical scale of $H = \epsilon H_g = \epsilon L(e)^{1/2}$, where $\epsilon = f_0^2 N^2$, $N^2$ is the square of the Brunt–Väisälä frequency, $f_0$ is the reference Coriolis parameter, and $L$ is the characteristic horizontal length scale of the system.

FIG. 6. Geostationary Operational Environmental Satellite (GOES-East) 0.6-µm water vapor channel satellite imagery for (a) 0000 UTC 24 Sep, (b) 0000 UTC 25 Sep, (c) 0000 UTC 26 Sep, and (d) 0000 UTC 27 Sep 2003.
vortex in a highly convective environment is common as models try to “lock on” to the correct incipient circulation.

By 0000 UTC 25 September, a pronounced closed low-level circulation (L) becomes evident northeast of the upper-level trough (Fig. 9c). While this feature appears to represent Juan in the EDAS, it is difficult to confirm this assertion, given a continued discrepancy between the location of this feature and Juan’s best-track location (Fig. 1). The cyclonic vortex (L) is better defined by 1200 UTC 25 September, centered near 29°N, 62°W (Fig. 9d). The scale of the vortex is large (∼1000 km) and almost synoptic in nature at this time. Only 12 h later (0000 UTC 26 September; Fig. 9e) does the scale of the vortex collapse to a size consistent with the compact Hurricane Juan (now labeled “J” in Fig. 9e). The evolution of the wind field here poses an interesting question. Did Juan develop from a concentration of vorticity by convection in the presence of an almost synoptic-scale low pressure system, or did Juan develop separately with its circulation disrupting the larger cyclone? This question is difficult, if not impossible, to answer with the available data. However, it is clear that Juan develops in a region conducive to cyclogenesis under the weak extratropical cyclone TT paradigm of Davis and Bosart (2004). By 1200 UTC 26 September (Fig. 9f), Juan’s circulation (J) has strengthened appreciably and is finally collocated with the NHC best-track position (Fig. 1). Note that even at this time, Juan is located in a region of local maximum in $P_{DT}$ and minimum of the coupling potential.

The nature of Juan’s genesis can be further explored by an examination of the Eady-type maps (Eady 1949) shown in Fig. 10. These plots display the $\theta_{DT}$ anomaly $\theta_{DT}$ relative to a balanced mean state (described below) and the mean sea level pressure. While temperature perturbations on the lower boundary are more commonly displayed in the classic Eady model of develop-
ment, pressure can be substituted under the constraints of the model since lower-boundary temperature perturbations must hydrostatically be reflected in the height field at the bottom boundary given the restriction of null interior PV. The basic state (denoted by overbars in the following equations) consists of an atmosphere with a constant buoyancy forcing,

$$N^2 = -\frac{g}{\rho} \frac{\partial \rho}{\partial z}.$$  

(1)
Fig. 9. Dynamic tropopause pressure (solid lines plotted at 50-hPa intervals) and coupling potential in shading with values as given by the attached grayscale. The 925–850-hPa layer-averaged winds are plotted the same as in Fig. 5. The DT is here taken to be the 2-PVU surface. Panels are plotted at 12-h intervals beginning at 0000 UTC 24 Sep 2003 and correspond to (a) 0000 UTC 24 Sep, (b) 1200 UTC 24 Sep, (c) 0000 UTC 25 Sep, (d) 1200 UTC 25 Sep, (e) 0000 UTC 26 Sep, and (f) 1200 UTC 26 Sep 2003. Details concerning specific annotations are contained in the text.
a constant vertical wind shear, 
\[ \frac{\partial U}{\partial z} = U_0, \]  
(2) 
a purely meridional temperature gradient, 
\[ \nabla \cdot \overline{T} = \frac{\partial T}{\partial y} = B, \]  
(3)

FIG. 10. Dynamic tropopause $\theta$ anomalies (in K with values as shown on the attached color bar) and reduced mean sea level pressure (contour interval of 4 hPa). The DT is here taken to be the 2-PVU surface and anomalies are computed relative to Eady model basic state as described in section 4c of the text. Panels are plotted at 12-h intervals beginning at 0000 UTC 24 Sep 2003 and correspond to (a) 0000 UTC 24 Sep, (b) 1200 UTC 24 Sep, (c) 0000 UTC 25 Sep, (d) 1200 UTC 25 Sep, (e) 0000 UTC 26 Sep, and (f) 1200 UTC 26 Sep 2003. Details concerning specific annotations are contained in the text.
and no latitudinal variation in wind speed. Here $\rho$ represents the air density, $U$ is the zonal component of the wind, $T$ is the dry bulb temperature, and $g$ is the gravitational constant. The $\nabla_H$ operator acts in two dimensions on a constant height surface. On an $f$ plane, this basic state clearly contains no PV gradient,

$$-\frac{1}{\rho} (\zeta + f) \frac{\partial \theta}{\partial z} = C,$$

where C is a constant that depends on the basic-state thermal structure and reference latitude. Based on the International Civil Aviation Organization Standard Atmosphere and mean September conditions, values for the free parameters of the Eady background are $N^2 = 0.0128$ s$^{-2}$ (the square of the Brunt–Väisälä frequency) and $B = 6 \times 10^{-6}$ K m$^{-1}$ on an $f$ plane centered at 45°N. Surface and tropopause $\theta$ perturbations from this basic state are more representative of Eady model forcings than are departures from climatological means since stationary forcings are known to exist in the time-averaged fields. Although the classic normal mode instability mechanism in the Eady model controls long-term development, the transient superposition of upper- and lower-boundary anomalies can have a strong influence on intensity during the early stages of the system’s evolution. If the anomalies are copropagating and in resonance, then this “unshielding” of the wind fields associated with them may lead to significant development in the absence of normal mode instability (de Vries and Opsteegh 2005). It is this superpositional growth habit that will be of primary interest here.

At 0000 UTC 24 September, cold (negative) $\theta_D^{\Delta T}$ values of $-20$ to $-25$ K are prevalent over the western North Atlantic south of 30°N (Fig. 10a). These cold anomalies are generally collocated with a surface low pressure system with a steady minimum central pressure of 1013 hPa (labeled “L” in Fig. 10). Between 1200 UTC 24 September and 0000 UTC 25 September (Figs. 10b,c), a cold (trough) anomaly (feature “A” in Fig. 10) moves off the eastern seaboard and toward the persistent Caribbean cold anomaly (labeled “T”) and the surface low (L) shrinks and deepens slightly to 1011 hPa (Fig. 10c). As the incipient storm (L) moves north in response to the approaching trough (A), a weak region of warm $\theta_D^{\Delta T}$ starts to develop to the north of the center (Fig. 10d). Once again, it is difficult to assess the precise location of Juan in the analysis since the scale of the surface low is considerably bigger than that of Juan. By 0000 UTC 26 September (Fig. 10e), a region of warming is evident on the DT north and east (downshear) of Juan (now identified by “J”), presumably in response to the convection associated with the hurricane (see WV imagery, Fig. 6c). As a result of this warming, a sharp gradient $\theta_D^{\Delta T}$ develops with a magnitude of about 25 K (200 km)$^{-1}$. Following 0000 UTC 26 September there is a collapse in the scale of the sea level pressure minimum, raising the question of whether this vortex (J) is simply an evolution of the persistent low pressure system (L) or the result of a secondary spinup as discussed in the description of the multiple lower-level vorticity features present in Fig. 5j (section 4b). In either case, the development of Juan continues from this point, with a drop in the minimum central pressure to 1008 hPa by 1200 UTC 26 September (Fig. 10f). Once again, the development occurs on in a region of $\theta_D^{\Delta T}$ gradient, suggestive of baroclinically induced cyclogenesis.

### d. A quasigeostrophic perspective

In this section, the synoptic-scale mechanisms modulating the vertical motion associated with the development of Juan are analyzed. From a QG perspective, forcing for ascent can be attributed to differential vorticity advection and the horizontal Laplacian of temperature advection in the absence of diabatic and frictional effects as demonstrated, for example, by Trenberth (1978). Alternatively, this can be expressed as the advection of geostrophic absolute vorticity by the thermal wind (Sutcliffe 1947; Sutcliffe and Forsdyke 1950). While diabatic effects are certainly at work in this case—as evidenced by local nonconservation of $\theta_D^{\Delta T}$ in Fig. 5—the presence of convection does not render QG diagnostics ineffective. On the contrary, the fact that convectively driven vertical motion is implicitly ignored under the QG framework allows for a clean partitioning of the ascent forcing into large-scale environmentally forced ascent and locally generated convective components, the former of which is considered explicitly here. Additionally the QG perspective is being adopted as a diagnostic, and not as a predictive tool.

Figure 11 is a simple representation of the Sutcliffe approximation and displays the 1000–200-hPa thickness and the 700–400-hPa layer-averaged geostrophic absolute vorticity. At 0000 UTC 24 September, a weak trough in the thickness field is located near 79°W as indicated by the dotted trough line in Fig. 11a. A long, narrow region of vorticity stretches from 73° to 50°W near 27°N, labeled with a dashed line in the same figure (Fig. 11a only). The variable thermal wind “blowing” across this vorticity band at various angles suggests that scattered areas of forcing for vertical motion exist near localized regions of vorticity gradients surrounding the line. This is consistent with the scattered areas of con-
Fig. 11. Layer-mean absolute vorticity (700–400 hPa, with shading as indicated in the grayscale and a scaling factor of $10^{-5}$ s$^{-1}$) and 1000–200-hPa thickness (solid lines plotted at 6-dam intervals). Panels are plotted at 12-h intervals beginning at 0000 UTC 24 Sep 2003 and correspond to (a) 0000 UTC 24 Sep, (b) 1200 UTC 24 Sep, (c) 0000 UTC 25 Sep, (d) 1200 UTC 25 Sep, (e) 0000 UTC 26 Sep, and (f) 1200 UTC 26 Sep 2003. Details concerning specific annotations are contained in the text.
vection visible on the water vapor imagery at this time (Fig. 6a). This pattern persists over the next 12 h (Fig. 11b), changing only by 0000 UTC 25 September as a thermal ridge develops south of 30°N at about 63°W (Fig. 11c) in response to the convection associated with Juan. The ridge reorients the thermal wind in the upstream region to southwesterly, thereby increasing the implied QG forcing for ascent as it blows at a greater angle across the vorticity gradient. This results in a redistribution of the ascent to the northeast of the vorticity center. As the thermal trough (located near 72°W) moves eastward, the thickness gradient between the trough and the ridge strengthens, resulting in an increase in the magnitude of the thermal wind at 1200 UTC 25 September (Fig. 11d). It is at this point that the QG forcing for ascent is at its strongest, and is consistent with the rapid intensification of Juan in the analysis fields at this time. Only 12 h later (0000 UTC 26 September; Fig. 11e), the circulation of Juan is collocated with a thickness ridge, indicating a more classic tropical cyclone structure and representing an effective end to any significant QG forcing for ascent.

As discussed briefly in section 2, the EDAS and GDAS analyses portray different near-storm lower-level vorticity structures during Hurricane Juan’s initiation stage. Figure 12 shows a direct comparison of the QG diagnostics of the two analyses for 0000 UTC 25 September (cf. Figs. 5i and 8b) and provides a strong motivation for the use of the high-resolution EDAS product. The vorticity band described above appears as a coherent feature east of Florida at 27°N, 65°W in the EDAS analysis (Fig. 12a) but is reduced to a weak, amorphous structure in the GDAS fields (Fig. 12b). A set of ship observations shown in both panels of Fig. 12 suggest a sharp northeast–southwest oriented shear line consistent with a surface reflection of the EDAS vorticity pattern. The implied QG ascent forcing during the reorientation or restructuring of this band is much stronger than it is for the thermal wind advection of the largely uniform vorticity structure in the GDAS analysis. These results suggest that the EDAS’s ability to resolve the verifying vorticity band along which the incipient vortex forms is necessary for a complete understanding of important QG genesis mechanisms at work in this case.

5. Summary and discussion

This study has focused on the analysis of high-latitude landfalling Hurricane Juan (in 2003) and the midlatitude environment into which the storm propagated. A diverse set of analysis methods have been used to diagnose important components of Juan’s complex life cycle. These results suggest that high-amplitude prestorm ridging plays an essential role in the storm’s evolution by creating an environment both dynamically and thermodynamically favorable for the maintenance of tropical structures at high latitudes. To evaluate the generality of this finding, a basic compositing study has been presented for a set of TCs broadly similar to Juan. A strong east coast trough–ridge couplet has thereby been found to be both a statistically and a physically significant feature for high-latitude landfalling hurricanes in the Atlantic basin.

The EW precursor for Hurricane Juan can be traced back to the West African coast in the middle of September 2003. However, it was not until this wave interacted with a shear line off the North American east coast that convection became effectively focused and development began. The incipient vortex grew in a baroclinic environment beneath the tropopause depression associated with the deep low-latitude front. Modified Eady (Eady 1949) model diagnostics suggest that the interaction between this upper-level feature and the developing lower-level structure triggers a superposition growth mechanism that provides an impetus for accelerated development. Similar features have been found to be important to the TT process in the investigations of Diana (in 1984; Bosart and Bartlo 1991) and Michael (in 2000; Davis and Bosart 2003) and follow the weak extratropical cyclone paradigm proposed by Davis and Bosart (2004). Even following the development of a warm core perturbation, links to Juan’s baroclinic origins persisted at upper levels. Further interaction with the midlatitude flow occurred during Juan’s track toward Nova Scotia as a shortwave perturbation originating in the subtropical jet–induced oscillations in the vortex. After landfall, Hurricane Juan weakened slowly as it traversed the Maritimes and began ET before being consumed by a broad midlatitude trough.

A composite study of eight northward-moving, high-latitude landfalling hurricanes between 1948 and 2002 has been presented in order to highlight generally important features in the midlatitude flow. The persistence of the midlevel east coast ridging over the period leading up to landfall suggests that this important feature has midlatitude origins and is not simply the result of the hurricane’s outflow. Rossby wave trains appear to have played a role in the development of the ridge; however, further analysis of the dynamics and energetics of the midlatitude flow (and the possibility of diabatic Rossby wave generation by the composite hurricane) will be left for a future study.

The dynamical and statistical significance of the
strong east coast ridge–trough couplet, and its role in developing a strong meridional flow over the continent, represents an important finding of this study. The mid-latitude environment into which Hurricane Juan was steered was both thermodynamically and dynamically favorable for the maintenance of the hurricane vortex. Elevated lower-level \( \theta_v \) values and low coupling indices reduced the bulk potential stability of the hurricane’s surroundings while minimal vertical shear did little to tilt the vortex or induce secondary circulations until the storm accelerated toward Nova Scotia. The strength and persistence of the east coast trough–ridge couplet both in the case study and in the composite analysis suggest that this feature may prove to be a key element

![Diagram](image-url)
in the understanding and forecasting of high-latitude landfalling hurricanes.

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