The Role of Warm Conveyor Belts for the Intensification of Extratropical Cyclones in Northern Hemisphere Winter

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ABSTRACT

The role of warm conveyor belts (WCBs) and their associated positive low-level potential vorticity (PV) anomalies are investigated for extratropical cyclones in Northern Hemisphere winter, using ERA-Interim and composite techniques. The Spearman correlation coefficient of 0.68 implies a moderate to strong correlation between cyclone intensification and WCB strength. Hereby, cyclone intensification is quantified by the normalized maximum 24-h central sea level pressure deepening and WCB strength by the WCB air mass associated with the cyclone's 24-h period of strongest deepening. Explosively intensifying cyclones typically have strong WCBs and pronounced WCB-related PV production in the cyclone center; they are associated with a WCB of type W2, which ascends close to the cyclone center. Cyclones with similar WCB strength but weak intensification are either diabatic Rossby waves, which do not interact with an upper-level disturbance, or cyclones where much of the WCB-related PV production occurs far from the cyclone center and thereby does not contribute strongly to cyclone deepening (WCB of type W1, which ascends mainly along the cold front). The category of explosively intensifying cyclones with weak WCBs is inhomogeneous but often characterized by a very low tropopause or latent heating independent of WCBs. These findings reveal that (i) diabatic PV production in WCBs is essential for the intensification of many explosive cyclones, (ii) the importance of WCBs for cyclone development strongly depends on the location of the PV production relative to the cyclone center, and (iii) a minority of explosive cyclones is not associated with WCBs.

1. Introduction

Many studies have shown that moist diabatic processes play an important role for the evolution of extratropical cyclones [see, e.g., the review by Uccellini (1990)]. Particularly, several case studies have revealed a significant contribution of latent heating to the intensification of explosively deepening cyclones (e.g., Reed et al. 1992; Wernli et al. 2002; Ludwig et al. 2014). Theoretical studies confirmed the importance of latent heating for cyclone development, revealing an increased intensification rate and structural changes in the presence of moisture (e.g., Mak 1982; Emanuel et al. 1987). The majority of these

studies adopted a potential vorticity (PV) perspective (Hoskins et al. 1985), where cyclone development is described by the interaction of an upper-level positive PV anomaly, a surface thermal anomaly, and a diabatically produced low-level positive PV anomaly. In the mature stage of a cyclone, the three anomalies can become vertically aligned to form a so-called PV tower—a column of anomalously high PV values spanning the troposphere and inducing a strong cyclonic circulation (Hoskins 1990; Rossa et al. 2000; Badger and Hoskins 2001). The diabatically produced low-level PV anomaly can significantly enhance the low-level circulation in cyclones and thereby contribute to their intensification (Davis and Emanuel 1991; Stoelinga 1996). It is also known that intense diabatic heating and low-level PV production in extratropical cyclones occurs in the warm conveyor belt (WCB; e.g., Wernli and Davies 1997). In this study we investigate whether these diabatically produced positive PV anomalies in WCBs affect cyclone intensification—that is, whether they become involved in the formation of PV towers in the center of cyclones. The following paragraphs provide more

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detailed information on theoretical studies on moist baroclinic instability, diabatic processes in real cyclones, and WCBs.

a. Moist baroclinic instability

Theoretical studies of moist baroclinic waves using linear and nonlinear models and different latent heating parameterizations have demonstrated that diabatic processes increase the growth rate and lead to a contraction of the region of ascent (Mak 1982; Emanuel et al. 1987; Whitaker and Davis 1994). An increase in moisture results in an enhanced cyclone intensification rate, a decrease in the minimum central pressure, and an increase in precipitation and extreme surface winds (Booth et al. 2013).

De Vries et al. (2010) developed a theoretical framework to describe the evolution of baroclinic waves in the presence of moisture. In this framework diabatic PV production is parameterized in terms of vertical motion, and PV anomalies are linearly partitioned into dry and moist components, which can interact with each other by their induced meridional and vertical velocities. Four unstable interactions have been identified between moist and dry waves in different layers, ranging from the classical dry instability (Eady 1949) to fully moist instability, with two intermediate types resulting from the interaction between a moist and a dry wave [see Cohen and Boos (2016) for a detailed review]. Hence, in three out of the four categories diabatic processes are crucial, but their interaction with the rest of the dynamics is different in each case.

b. Diabatic processes in real cyclones

The importance of moist processes for cyclone intensification has been pointed out in numerous case studies. Examples of strongly diabatically influenced, explosively intensifying cyclones include the Queen Elisabeth II storm in 1978 (Anthes et al. 1983; Gyakum 1983a,b), the Presidents' Day storm in 1979 (Bosart 1981; Atlas 1987; Whitaker et al. 1988), the Scamp storm in 1987 (Reed et al. 1992; Stoelinga 1996), Lothar in 1999 (Wernli et al. 2002; Rivière et al. 2010), and, more recently, Xynthia in 2010 (Fink et al. 2012; Liberato et al. 2013; Ludwig et al. 2014). In their mature stage, most of these cyclones were associated with a pronounced PV tower, with the low-level positive PV anomaly resulting from intense condensational PV production. In a climatological study, Čampa and Wernli (2012) showed that the amplitude of the upper-level, low-level, and surface positive PV anomalies tend to be stronger in intense than in weak cyclones. In the eastern North Atlantic, Dacre and Gray (2013) found a significant sensitivity of cyclone intensity after 48 h to the presence of a

diabatically produced midlevel PV anomaly at the initial time. They suggested the PV anomaly to be related to diabatic processes within the WCB—however, without explicitly identifying the WCB by means of strongly ascending trajectories.

Two special categories of cyclones with an important contribution from low-level latent heating are type C cyclones (Deveson et al. 2002; Plant et al. 2003; Ahmadi-Givi et al. 2004) and diabatic Rossby waves (DRWs; Parker and Thorpe 1995; Boettcher and Wernli 2011, 2013). Type C cyclones are characterized by an interaction between an upper-level dry and a diabatically generated low-level PV anomaly, while the surface thermal anomaly plays a minor role for development. DRWs are low-level positive PV anomalies located in a moist, baroclinic environment, sufficiently distant from an upper-level disturbance such that coupling does not occur. They propagate rapidly along the baroclinic zone by continuous regeneration of the DRW downstream of its original position through diabatic processes.

In addition to the role of latent heating for cyclone intensification, a few publications suggested that moist processes can also influence the track of extratropical cyclones. In idealized cyclones, Coronel et al. (2015) found a faster northeastward and cross-jet motion in the presence of moisture, corroborating earlier findings of idealized studies (Mak 1982) and simulations of real cyclones (e.g., Davis et al. 1993). They explained the eastward motion by the diabatically produced low-level PV anomaly as in the case of DRWs and the northward motion by nonlinear advection by the upper-level positive and diabatically enhanced negative PV anomalies forming up- and downstream, respectively. The importance of the upper-level PV and diabatic heating for the cyclone's poleward propagation was confirmed by Tamarin and Kaspi (2016), who performed a detailed PV tendency analysis for cyclones in an idealized GCM.

c. Warm conveyor belts

The WCB plays a key role when investigating diabatic processes from a Lagrangian perspective. It is a coherent warm and moist airstream, which originates in the boundary layer of the cyclone's warm sector and ascends rapidly to the upper troposphere while moving toward polar regions (e.g., Harrold 1973; Browning et al. 1973; Carlson 1980). The WCB is the primary cloud- and precipitation-generating flow in extratropical cyclones (Browning 1990), and it plays a crucial role for total and extreme precipitation events in many regions of the extratropics (Pfahl et al. 2014). In a detailed case study on microphysical processes along a WCB, Joos and Wernli (2012) showed that several microphysical processes can modify PV, in particular condensation of water vapor at

low levels, and depositional growth of snow at upper levels. In the Northern Hemisphere, WCBs are more frequent in winter than in summer, with two preferential ascending regions in the western North Atlantic and North Pacific (Eckhardt et al. 2004; Madonna et al. 2014).

In many descriptions the WCB is divided into two branches (Young et al. 1987; Browning and Roberts 1994; Browning 1999; Martínez-Alvarado et al. 2014). The primary branch, W1, travels along the cold front, and depending on the movement of the air relative to the cold front it can be distinguished between rearward-sloping ascent above the cold front and forward-sloping ascent above the warm front (cf. Browning 1990). The secondary branch, W2, approaches the cyclone center and rises above the bent-back front into the cloud head. In a case study of an extratropical cyclone, Wernli (1997) identified both W1 WCB trajectories rising at the cold and warm fronts, respectively, and a W2 flow ascending in the close vicinity of the cyclone center. A WCB of type W2 has also been identified in idealized experiments of moist baroclinic waves (Schemm et al. 2013).

As mentioned above, the intense WCB-related cloud–diabatic processes produce a strong positive PV anomaly in the lower and middle troposphere. At upper levels diabatic PV destruction leads to low PV values in the WCB outflow. These produce significant negative PV anomalies in the tropopause region, which can interact with the extratropical waveguide and thereby substantially influence the downstream flow (e.g., Wernli 1997; Pomroy and Thorpe 2000; Massacand et al. 2001; Ahmadi-Givi et al. 2004; Grams et al. 2011; Methven 2015).

In this study we focus on the low-level positive PV anomaly that occurs during the early phase of the WCB ascent, and its influence on the associated cyclone. Very little is known about the location of strongest WCB-related ascent and PV production relative to the cyclone center and whether these positive PV anomalies contribute to cyclone intensification. In addition, no studies are available on the relative importance of the W1 and W2 types of WCBs for cyclone development.

d. Objectives and structure of this study

The goal of this study is to quantify the role of WCBs and their associated positive PV anomalies for the intensification of a large set of cyclones in Northern Hemisphere (NH) winter. To this end, cyclone intensification will be measured as the normalized maximum sea level pressure (SLP) deepening over 24 h (Sanders and Gyakum 1980), and WCB strength will be quantified as the number and mass of the WCB trajectories associated with the cyclone at low levels during its strongest intensification (see sections 2d and 2e). An

illustrative case study and detailed climatological analyses will be used to address the following questions:

- 1) For a climatological set of cyclones, is there a correlation between cyclone intensification and the strength of the associated WCB?
- 2) What is the structure and evolution of strongly deepening cyclones with an intense WCB, and where do the WCB-related positive PV anomalies occur relative to the cyclone center?
- 3) How do strongly deepening cyclones with an intense WCB differ in terms of structure and evolution from (i) weakly deepening cyclones with an intense WCB and (ii) strongly deepening cyclones with a weak WCB?

In the supplemental material it is assessed how sensitive the answers to these questions are for different measures of cyclone intensification. For the climatological analyses, we will use composite techniques, which have proven to be insightful to examine the common characteristics of particular types of cyclones (e.g., Field and Wood 2007; Bengtsson et al. 2009; Catto et al. 2010; Dacre et al. 2012).

The structure of the remainder of this paper is as follows. Section 2 describes the methods used in this study. In section 3a the statistical connection between cyclone intensification and WCB strength is discussed, and based on these measures three different cyclone categories are defined. Section 3b presents the spatial distribution of the three categories. The category of explosively intensifying cyclones with a strong WCB is discussed in section 4, the category of weak cyclones with a strong WCB in section 5, and the category of explosively intensifying cyclones with a weak WCB in section 6. Section 7 provides a summary of the results and the main conclusions.

2. Data and methods

a. Data

This study is based on the ERA-Interim dataset from the European Centre for Medium-Range Weather Forecasts (Simmons et al. 2007; Dee et al. 2011). The analysis is confined to the NH winter (December–February) for the period 1979–2014. ERA-Interim data are available at a spectral resolution of T255 on 60 vertical levels and at a temporal resolution of 6 h. The original fields are interpolated onto a $1^\circ \times 1^\circ$ longitude–latitude grid. From the primary data, potential vorticity, potential temperature θ , the horizontal wind velocity, and the upper-level forced quasigeostrophic (QG) vertical motion are calculated as secondary variables. The QG vertical motion is calculated with the diagnostic of Clough et al. (1996). As in Boettcher

and Wernli (2011, 2013), the vertical motion induced by the upper levels, defined as the layer between 500 and 100 hPa, is evaluated at 700 hPa.

b. WCB identification

The study makes use of the WCB climatology provided by Madonna et al. (2014). It is based on trajectory calculations with the Lagrangian Analysis Tool (LAGRANTO; Wernli and Davies 1997; Sprenger and Wernli 2015) and the application of a Lagrangian selection criterion that accounts for the strong ascent of WCBs (Wernli and Davies 1997). Forward trajectories are started globally, every 6 h from an equidistant grid with $\Delta x = 80$ -km horizontal and $\Delta p = 20$ -hPa vertical resolution in the lower troposphere between 1050 and 790 hPa. WCB trajectories are then identified as air parcels that ascend more than 600 hPa within 48 h from the boundary layer to the upper troposphere. Furthermore, in order to exclude deep convective systems, the horizontal position of the WCB trajectory must be collocated with a surface cyclone for at least one 6-hourly time step during the ascent [see Madonna et al. (2014) for details]. The trajectory data consist of the horizontal position, pressure, and the evolution of different physical parameters along the WCB path every 6 h during the 2-day ascent phase.

c. Cyclone identification and tracking

The climatology of extratropical cyclones is based on a slightly modified version of the procedure developed by Wernli and Schwierz (2006); the same as used in Madonna et al. (2014). It consists of an identification algorithm, which yields cyclone areas every 6 h, and a separate tracking algorithm. A surface cyclone is identified as the two-dimensional area surrounding one or several local SLP minima enclosed by the outermost closed SLP contour, with the maximum length of this contour restricted to 7500 km. The enclosed area is regarded as the cyclone area. The tracking algorithm connects cyclone centers by searching the most likely continuation for each SLP minimum among the identified minima 6 h later.

While the tracking algorithm mostly performs well, in some cases long-lived systems are erroneously split into shorter-lived systems. In addition to very short tracks, this can result in unrealistic cyclone life cycles—for instance, cyclones with a very low SLP minimum and a large area already at the beginning of their tracks. To retain only tracks with a realistic and well-defined cyclone life cycle, in the current analysis it is specified that (i) the track duration is at least 48 h and (ii) at the first time step of the track the pressure difference between the outermost closed SLP contour and the SLP minimum is less than 5 hPa. This threshold value of 5 hPa has

been subjectively chosen by analyzing a number of critical tracks. Of course not all unrealistic cases can be filtered out with these requirements, while others are erroneously excluded, but nevertheless a detailed investigation of many cases showed that the large majority of the retained tracks has a well-defined and realistic cyclone life cycle.

In addition to the two filters mentioned above, tropical cyclones are excluded from the analysis by considering only tracks located north of 25°N at least for one 24-h interval. In total, over the 36-yr study period 5069 extratropical cyclones are found that meet these criteria.

d. Attribution of WCBs to cyclones

To investigate the interaction between cyclone intensification and WCBs, every WCB trajectory is attributed to a cyclone. As described above, in our climatology it is requested that every WCB trajectory matches with a surface cyclone area at least once during the 2-day ascent. Here use is made of this criterion by identifying for every cyclone, during the entire cyclone life cycle, all WCB trajectories that overlap with the horizontal position of the cyclone area. Not all cyclones are associated with WCB trajectories, but—because of the matching criterion set by Madonna et al. (2014)—every WCB trajectory can be attributed to a cyclone.

In some cases a WCB trajectory matches with more than one cyclone during the 48 h. Such a double counting can occur, for instance, when a WCB trajectory matches with a first cyclone during the ascent, and with a second cyclone when the WCB outflow later moves over the area encompassed by another cyclone. Since the focus of this study is on the role of the low-level PV produced during the WCB ascent, in cases of several matches at different stages of the 2-day ascent, a WCB trajectory is attributed only to the first cyclone. By assigning every WCB trajectory to one cyclone track, we obtain for every cyclone the number of associated WCB trajectories, as well as their position and physical properties along the entire cyclone life cycle.

e. Measures for cyclone intensification and WCB strength

Following the approach of Sanders and Gyakum (1980), cyclone intensification is quantified in units of Bergeron, with

$$\Delta \text{SLP}_B = \frac{\Delta \text{SLP}}{24} \frac{\sin(60^\circ)}{\sin(\phi)}.$$

Here, ΔSLP is the change of the minimum SLP (hPa) over a 24-h interval along the track and ϕ is the mean latitude of the cyclone center during this interval. The

cyclone deepening rate ($\Delta\text{SLP}_{B,\text{max}}$) is then defined as the maximum value of ΔSLP_B of all 24-h intervals along the track. In the present study, only 24-h intervals are considered in which the cyclone is located entirely poleward of 25° latitude. After Sanders and Gyakum (1980), cyclones with a deepening rate of at least 1 Bergeron are considered to be intensifying explosively as so-called bombs. This corresponds to a deepening rate of 24 hPa in 24 h at 60°N . Alternative cyclone intensification measures that take into account the climatological pressure distribution or are based on relative vorticity, and their relationship with WCBs, are discussed in the supplemental material.

The WCB strength associated with a certain cyclone is measured by counting the number of WCB trajectories that are attributed to this cyclone and located at low levels (pressure > 500 hPa) sometime during the 24-h interval of strongest intensification. The trajectories can in principle be outside the cyclone area during this 24-h period, as long as they match with the cyclone at least once during their 2-day ascent. Because of the equidistant starting grid of the WCB trajectories (see section 2b) every WCB trajectory represents the same amount of air mass [i.e., $\Delta m \approx (1/g)(\Delta x)^2 \Delta p \approx 1.3 \times 10^{12}$ kg, with $g = 9.81 \text{ m s}^{-2}$], and the total WCB air mass at low levels associated with a cyclone is obtained by multiplying the number of WCB trajectories by Δm .

To evaluate the statistical relationship between cyclone intensification and WCB strength, the Spearman rank correlation coefficient is used (Spearman 1904). It is a nonparametric measure for the monotonic relationship between two variables.

3. Connection between cyclone intensification and WCB strength

a. Statistical relationship between cyclone intensification and WCB strength

Figure 1 shows for all ~ 5000 cyclone tracks a two-dimensional histogram of the statistical relationship between the WCB strength and the cyclone deepening rate ($\Delta\text{SLP}_{B,\text{max}}$) and, separately, their one-dimensional histograms. The distributions of WCB strength and cyclone intensification are positively skewed toward intense WCBs and strong cyclone intensification rate, respectively. The latter is in accordance with the statistical analyses by Roebber (1984) and Gyakum et al. (1989).

The red and light green colors in the two-dimensional histogram reveal that many cyclones are associated with a weak deepening and few WCB trajectories. Effectively, 45% of all cyclones are not associated with any WCB trajectory (of which 98.5% have a deepening rate of less

than 1 Bergeron), which is in good qualitative agreement with the findings of Eckhardt et al. (2004). They mainly occur in regions of low climatological WCB frequency (Eckhardt et al. 2004; Madonna et al. 2014), in particular at high latitudes, in the Mediterranean region and over the continents (not shown). The dark green and blue colors in Fig. 1 indicate a tendency that more intense cyclones are associated with stronger WCBs. This is confirmed by the Spearman rank correlation coefficient of 0.68, which implies a moderate to strong positive correlation between cyclone intensification and the strength of the associated WCB. In numbers, the average deepening rate increases from 0.28 Bergeron for cyclones with less than 50 WCB trajectories, to 0.8 Bergeron for cyclones with a WCB trajectory number between 50 and 1500, and to 1.35 Bergeron for cyclones with more than 1500 WCB trajectories. However, a large scatter is apparent with some cyclones intensifying strongly without any WCB contribution, and others having a strong WCB but intensifying only weakly.

The position of some prominent winter storms is marked by letters in the phase space in Fig. 1. They all intensified explosively, and most of them were associated with a strong WCB. Among those with a particularly strong WCB are the storms Martin (Ulbrich et al. 2001) and Xynthia (Liberato et al. 2013), which caused major damage in western and central Europe, as well as several memorable storms along and off the U.S. East Coast, like the Presidents' Day snowstorm (Bosart 1981; Atlas 1987; Whitaker et al. 1988), the snowstorms of 4–5 February 1995 (Nicosia and Grumm 1999) and 24–25 January 2000 (Zhang et al. 2002), and the devastating windstorm Scamp (Reed et al. 1992; Stoelinga 1996).

The winter storms Lothar and Klaus had a weaker WCB, although they have been identified as strongly diabatically driven cyclones (Wernli et al. 2002; Fink et al. 2012). In comparison to most of the above-mentioned storms, which reached a considerable size during explosive intensification, Lothar and Klaus both remained relatively small-scale storms, and their weak WCB strength is therefore related to the small size of these cyclones. As a test, we performed an additional analysis, where the WCB strength is measured by the number of WCB trajectories normalized by the cyclone area. Using this measure, Klaus and Lothar both attain WCB strength at the high end of the distribution of all cyclones. On the other hand, for large cyclones such a normalization resulted in a weak WCB strength when the WCB trajectories were concentrated in a small area around the cyclone center. However, except for these two subgroups the distribution of the cyclones in the phase-space diagram

FIG. 1. (bottom left) Two-dimensional histogram showing for every NH winter cyclone the deepening rate ($\Delta\text{SLP}_{B,\text{max}}$) (Bergeron; bin width = 0.1 Bergeron) and the associated WCB strength, expressed by the mass of the WCB trajectories at low levels integrated over the 24-h period of strongest pressure deepening (bin width $\approx 0.33 \times 10^{15}$ kg, corresponding to 250 WCB trajectories). Colors indicate the number of cyclones in a specific bin. The corresponding one-dimensional histograms are shown for (top) cyclone deepening and (bottom right) WCB strength. The total number of cyclones is 5069. The Spearman rank correlation coefficient between cyclone intensification and WCB strength is 0.68. The boundaries of category C1, C2, and C3 are marked by the black lines, and the letters indicate the position of specific cyclones: “C”—C1 case study discussed in section 4a; “K”—storm Klaus; “L”—storm Lothar; “M”—storm Martin; “N₁” and “N₂”—nor’easter snowstorms in February 1995 and January 2000, respectively; “P”—Presidents’ Day cyclone; “S”—Scamp storm; and “X”—storm Xynthia.

remained very similar as in Fig. 1, indicating that for most cyclones it is meaningful to measure WCB strength in absolute terms—that is, not normalized by the cyclone area.

To understand whether and how WCBs can influence cyclone intensity, in the following the structure and evolution of different cyclone categories, defined in the phase space shown in Fig. 1, is investigated in detail. The category boundaries of the C1 cyclones are

defined by choosing from the explosively intensifying cyclones (deepening rate of at least 1 Bergeron) the 500 cases with maximum WCB strength. This yields a lower threshold of 2.78×10^{15} kg for the WCB strength (equivalent to 2130 WCB trajectories). As evident in Fig. 1, many of the above discussed well-known storms are in category C1. C2 cyclones ($n = 101$) have a similar WCB strength as C1 (at least 2.78×10^{15} kg) but a weak deepening rate of less than

FIG. 2. Spatial distribution of cyclones at the beginning of their 24-h period of maximum intensification. C1 cyclones are shown by gray labels ($n = 500$), C2 cyclones are blue for subtype C2a ($n = 29$) and green for subtype C2b ($n = 72$), and C3 cyclones are red ($n = 154$).

0.8 Bergeron. Finally, C3 cyclones ($n = 154$) are defined as explosively intensifying cyclones with a weak WCB with a maximum of 0.33×10^{15} kg (equivalent to 250 WCB trajectories; Fig. 1). Because of the positive correlation between cyclone intensification and WCB strength, on average the C1 cyclones have a stronger WCB than C2 (on average 5.37×10^{15} kg compared to 4.50×10^{15} kg in C2). Likewise, the average Bergeron value is larger in C1 (1.67 Bergeron) than in C3 (1.27 Bergeron).

While for C1 and C3 the categories are analyzed as a whole, for C2 two subtypes are distinguished: C2a cyclones ($n = 29$) do not interact with a tropopause disturbance (defined by the 2-PVU¹ contour on 315 K) during the entire 24-h period of strongest intensification, while C2b ($n = 72$) are associated with at least a weak tropopause interaction. The assignment to the two subtypes is based on manual analysis of the individual cyclones. As shown in the supplemental material, the attribution of the cyclones to the three categories and their mean properties are fairly robust to the measure used to define cyclone intensification.

b. Spatial distribution of the cyclones in C1, C2, and C3

Figure 2 shows the spatial distribution of the cyclones in the different categories at the beginning of their strongest deepening (referred to in the following as $t = 0$ h, while the middle and the end of the 24-h period of strongest SLP deepening will be referred to as $t = 12$ h and $t = 24$ h, respectively). The majority of the C1 cyclones is located between 25° and 50° N (mean latitude of 36° N) in the western and central ocean basins in regions of high climatological WCB frequency [Fig. 4 in Madonna et al. (2014)].

The majority of the C2 cyclones also occurs in regions of high climatological WCB frequency (Fig. 2). The C2a

cases are located relatively far south (mean latitude $< 30^\circ$ N at $t = 0$ h), and the C2b cyclones are slightly farther poleward (mean latitude of 36° N, as for C1).

In contrast to C1 and C2, many C3 cyclones occur at high latitudes or over the continents (Fig. 2)—both regions of low climatological WCB frequency. Additionally, a large number of C3 cyclones are located in the western and central ocean basins, although at more poleward latitudes than C1 and C2 (the mean latitude of all C3 cyclones is 49° N). Given their widespread occurrence, it is unlikely that C3 cyclones share a common evolution pattern (see later).

In the following the three categories are analyzed with the help of composites that are centered at the position of the SLP minimum. C1 cyclones are discussed first and in greatest detail in the next section, and their composites are shown at the beginning, in the middle, and at the end of the 24-h period of strongest intensification. The spatial structure of the C2 and C3 cyclones does not change considerably during their intensification. Therefore, for these categories only the composites at $t = 12$ h are discussed. Their composites at $t = 0$ h and $t = 24$ h are shown in the supplemental material. Additionally, in the supplemental material the temporal evolution of various parameters is compared for the three categories.

To identify a potential issue with conflation of basin-specific cyclone dynamics when compositing both Atlantic and Pacific cyclones, in each category separate composites have been created for the two regions (not shown). They showed that the overall structure and the main characteristics of the cyclones are very similar in both regions. In addition, it has been examined whether there is an issue in using different sample sizes in C1 ($n = 500$) and C2a ($n = 29$) by comparing the composites of 29 randomly selected C1 cyclones (not shown) to the entire C1 category. The structure of the subsample is very similar to the entire category, which indicates that a relatively small sample

¹ Potential vorticity unit; 1 PVU = $10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ K kg}^{-1}$.

of 29 cyclones is sufficient to capture the main features of a category.

4. Category C1—Explosively intensifying cyclones with strong WCBs

Before proceeding to the C1 composite analysis, a case study is described that represents the main characteristics of the C1 cyclones.

a. Case study

During the period 6–11 January 1990, an intense extratropical cyclone with a strong WCB developed over the North Atlantic. The system underwent an explosive deepening of 44 hPa between 1800 UTC 6 January and 1800 UTC 7 January (corresponding to 2.4 Bergeron) and was associated with a WCB strength of 8.49×10^{15} kg (see Fig. 1, label “C”). Overall, the minimum SLP dropped by almost 80 hPa in the course of 54 h.

The cyclone formed in the western North Atlantic (near 35°N, 74°E) at 1200 UTC 6 January south of an intense baroclinic zone. The system reached a minimum SLP of 1006 hPa 6 h later (Fig. 3a). At upper levels, a weak disturbance is located northwest of the developing system (Fig. 3a). Northeast of the cyclone, 250-hPa wind speeds of more than 90 m s^{-1} are present (Fig. 3b).

A strong WCB is associated with the cyclone. The WCB inflow at 1800 UTC 6 January occurs near the cyclone center and to the east along the baroclinic zone (Fig. 4a). Figure 3a shows the position and mean PV values of all gridded WCB air parcels that are located at low levels at 1800 UTC 6 January—that is, not only of the trajectories starting at that time step (Fig. 4a), but also of all WCB air parcels associated with the cyclone that started ascending at earlier time steps and are still in the lower troposphere. The PV values in Fig. 3a denote mean values over all trajectories at that grid point located somewhere in the vertical column between 1000 and 500 hPa. Slightly to the west of the cyclone center, the values are highest (more than 0.7 PVU; Fig. 3a) and related to midlevel air parcels, as evident in the west–east-oriented vertical section across the cyclone center (Fig. 5a). Another weak positive PV anomaly and coinciding WCB air parcels are present in the cyclone center at low levels (Fig. 5a). To the east, WCB trajectories with relatively low PV values of less than 0.5 PVU occur along the baroclinic zone (Figs. 3a,b). They correspond to boundary layer trajectories (Fig. 5a) and, hence, to the WCB inflow seen in Fig. 4a. WCB intersections with the 315-K isentrope to the west of the cyclone center (Fig. 3a) belong at that time to midlevel rather than upper-level

air parcels (Fig. 5a), which is due to the southerly position of the cyclone and accordingly the low altitude of the 315-K isentrope.

As the cyclone propagates farther northeastward it strongly intensifies, reaching a minimum SLP of about 981 hPa 12 h later at 0600 UTC 7 January (Fig. 3c). The cyclone is now located at the leading edge of an upper-level trough (Fig. 3c) near the right entrance of the slightly weakened jet (Fig. 3d). At the surface, cold and warm fronts start to develop. In addition to the WCB trajectories that have started at earlier time steps (Fig. 4a), a new WCB inflow forms to the east of the cyclone (Fig. 4b). Low-level WCB air parcels are present in the cyclone center and along the fronts (Fig. 3c). Importantly, the highest mean PV of the WCB air parcels occurs near the cyclone center, with values of almost 2 PVU, pointing to strong WCB-induced diabatic PV production. In the same region, a pronounced low-level PV maximum is visible both on 850 hPa (Fig. 3d) and in the vertical section (Fig. 5b). A vortex with high PV values of more than 1 PVU has developed and extends from the surface up to about 450 hPa, with a maximum of more than 2 PVU around 850 hPa. At the location of the vortex, a large number of WCB trajectories with high PV (cf. Fig. 3c) occur, corroborating that the pronounced PV anomaly has been formed by diabatic processes in the ascending WCB. To the west of the low-level vortex, a deep trough is present (Fig. 5b), which constitutes a favorable condition for the continuing explosive intensification of the storm.

The infrared satellite image shows cold cloud tops in the warm sector in the region of strongest ascent of the WCB (Fig. 6). Farther to the west, the motion of the trajectories along the cold front, which took place at earlier time steps as seen in Fig. 4a, manifests itself in an almost west–east-oriented cloud band extending from the U.S. East Coast toward the cyclone center, accompanied by a gradual increase of cloud-top height, and, most likely, embedded convection.

Minimum SLP amounts to 962 hPa 12 h later (i.e., at the end of the explosive deepening) (Fig. 3e). The upper-level disturbance has developed into an intense LC2-type wave (Thorncroft et al. 1993) wrapping cyclonically around the low-PV air. WCB intersection points at upper levels suggest the presence of both cyclonically and anticyclonically turning air in the WCB outflow, coupled with strong ridge amplification. Intense low-level diabatic PV production by WCB trajectories still occurs along the fronts and in the cyclone center, with anomalously high mean PV values of more than 2 PVU slightly southwest of the cyclone center (Fig. 3e). The low-level PV anomaly has further intensified

FIG. 3. Case study for category C1. Synoptic situation at (a),(b) 1800 UTC 6 Jan 1990, corresponding to $t = 0$ h; (c),(d) 0600 UTC 7 Jan, corresponding to $t = 12$ h; (e),(f) 1800 UTC 7 Jan, corresponding to $t = 24$ h; and (g),(h) 1800 UTC 8 Jan. (left) Mean PV value of gridded WCB air parcels with pressure > 500 hPa (PVU; shading), SLP (hPa; black contours every 3 hPa), PV on 315 K (blue contours for 2, 3, and 4 PVU), and intersection positions of WCB trajectories with the isentropic layer between 312.5 and 317.5 K (asterisks). (right) PV (PVU; shading) and potential temperature (black contours every 3 K) at 850 hPa, and wind speed at 250 hPa (green contours for 50, 60, 70, 80, and 90 m s⁻¹). The black cross marks the position of the cyclone center.

FIG. 4. Two-day WCB trajectories (colored by pressure; hPa) starting at (a) 1800 UTC 6 Jan 1990, corresponding to $t = 0$ h, and (b) 0600 UTC 7 Jan, corresponding to $t = 12$ h, and SLP (contours every 3 hPa) and potential temperature at 850 hPa (gray dashed contours for $\theta = 280, 285, 290, 295, \text{ and } 300$ K) at the start time of the trajectories. The black cross marks the position of the cyclone center.

(Figs. 3f and 5c), and above the low-level anomaly a narrow stratospheric intrusion is penetrating into the troposphere. Together they form an almost vertically aligned PV tower that induces a strong cyclonic circulation throughout the troposphere.

Over the next 12 h the pressure decreases at a similar rate, with continuous low-level diabatic PV production by the ascending WCB. At 0600 UTC 8 January the storm reaches its maximum intensity in terms of relative vertical vorticity at 850 hPa (not shown), but the

FIG. 5. West-east vertical cross sections through the cyclone center at (a) 1800 UTC 6 Jan 1990, corresponding to $t = 0$ h (from 77° to 57° W); (b) 0600 UTC 7 Jan, corresponding to $t = 12$ h (from 65° to 45° W); (c) 1800 UTC 7 Jan, corresponding to $t = 24$ h (from 51° to 31° W); and (d) 1800 UTC 8 Jan (from 36° to 16° W). In each case the cyclone center is located in the middle of the displayed cross section. Shown are PV (PVU; shading), potential temperature (contours every 5 K), and intersection positions of WCB trajectories with the cross section (black crosses).

FIG. 6. False-color infrared satellite image (brightness temperature; °C) at 0600 UTC 7 Jan 1990 derived from the GridSat-B1 data. Overlaid is the 2-PVU contour on 315 K (red) derived from ERA-Interim and the smoothed contours marking the grid points with at least one WCB trajectory with $p > 500$ hPa (black solid) and $p < 500$ hPa (black dashed).

strongest intensity in terms of SLP (936 hPa) is only reached 12 h later (Fig. 3g). At that time the cyclone is located below a deep tropopause (Fig. 5d) and associated with a T-bone frontal structure and an intense bent-back front (Fig. 3h; Shapiro and Keyser 1990). Weaker WCB-related PV production now only occurs along the warm front (Fig. 3g), and the PV tower is starting to decay (Fig. 5d).

In summary, this C1 case study suggests that the WCB played a significant role for cyclone development. The WCB-induced diabatic PV production in the cyclone center led to the formation of a strong low-level PV anomaly, which significantly enhanced the low-level circulation and thereby contributed to the explosive intensification of the cyclone.

b. Spatial structure of the C1 cyclones

1) STRUCTURE AT $t = 0$ h

Figure 7a shows the horizontal structure of the composite cyclone at $t = 0$ h in terms of SLP, PV on the 315-K isentrope, and the WCB trajectory frequency. The coordinates are defined relative to the position of the SLP minimum, which is relocated to 0° longitude and latitude, and the fields are averaged at these coordinates. Low-level and upper-level WCB trajectory frequencies are shown. The values are calculated such that they denote the percentage of C1 cyclones with at least one WCB air parcel with $p > 500$ hPa and $p < 500$ hPa,

respectively, at that grid point. For instance, a low-level trajectory frequency of 50% indicates that at that grid point 250 out of the 500 C1 cyclones are associated with at least one WCB trajectory somewhere between 1000 and 500 hPa. Additionally, the occurrence frequency of WCB trajectories with a high PV value of more than 1 PVU within the layer $900 < p < 700$ hPa is shown [similar to the high low-level PV (HLPV) WCB trajectories introduced by Madonna et al. (2014)] to get an impression of where the strongest WCB-related diabatic PV production occurs.

At $t = 0$ h the composite cyclone is located to the east of an upper-level trough (Fig. 7a) within a region of strong baroclinicity (not shown). Already at that early time step, a strong WCB is present in the cyclone's warm sector. The highest frequency of WCB trajectories at low levels is located slightly to the east of the cyclone center, with values of more than 70%, while the highest frequency of HLPV trajectories occurs directly in the cyclone center.² The WCB outflow into the downstream ridge is still relatively weak and has its

² Note that the centering of the composites on the SLP minimum favors the occurrence of high WCB trajectory frequencies near the center of the composite cyclone, while it more strongly smears out the signal farther away from the cyclone—for example, along the fronts. The high frequency of WCB trajectories at low levels in the cyclone center is nevertheless remarkable and consistent with the case study discussed earlier.

FIG. 7. Composites of C1 cyclones (a),(b) at the beginning ($t = 0$ h); (c),(d) in the middle ($t = 12$ h); and (e),(f) at the end ($t = 24$ h) of their strongest SLP deepening. (top) Frequency of low-level WCB trajectories with $p > 500$ hPa (%; shading; see text for details), of upper-level WCB trajectories with $p < 500$ hPa (blue contours at 5%, 10%, 20%, ..., 80%), and of WCB trajectories with $900 < p < 700$ hPa and $PV > 1$ PVU (green contours for 10%, 50%, and 65%), SLP (thin black contours every 3 hPa), PV on 315 K (thick black contours for 2, 3, and 4 PVU), and potential temperature on 850 hPa (black dashed contour for $\theta = 290$ K). (bottom) West-east vertical cross sections through the center of the composite cyclone (from -20° to 20°) showing PV (PVU; shading), potential temperature (contours every 5 K), and relative WCB trajectory density (%; red dashed for WCB trajectories with $p > 500$ hPa, blue dashed for WCB trajectories with $p < 500$ hPa, contours every 5%, and green solid for the 10% frequency of trajectories with $p > 500$ hPa and $PV > 1$ PVU).

maximum ($\sim 20\%$) slightly north of the low-level maximum.

The west-east vertical section across the center of the composite cyclone confirms the presence of a broad trough to the west and a ridge to the east of the cyclone (Fig. 7b). Furthermore, it shows a large relative WCB trajectory density³ between 1000 and 700 hPa near the cyclone center, with maximum values of about 50%. It is collocated with a positive PV anomaly, pointing to significant WCB-related diabatic PV production. Ahead of

the cyclone center, most trajectories are still in the boundary layer.

2) STRUCTURE AT $t = 12$ H

The SLP of the composite cyclone has dropped by more than 13 hPa to about 988 hPa 12 h later (Fig. 7c). The 850-hPa potential temperature field in Fig. 8a reveals the formation of surface cold and warm fronts, although they appear rather weak as a result of the compositing. The WCB has strengthened, and the area around the cyclone center associated with the most intense WCB-related PV production has increased (Fig. 7c). In the same region, high low-level PV (Fig. 8a) and a pronounced precipitation signal (Fig. 8b) point to strong latent heating by condensational processes. The diabatically produced PV anomaly extends from the surface up to about 550 hPa within an area of strong

³ The relative WCB trajectory density is defined as the WCB trajectory density normalized by the density of the WCB starting points, which is the maximum possible WCB trajectory density. For example, a WCB trajectory density of 50% in a certain air volume indicates that 50% of the air in that volume is WCB air.

FIG. 8. C1 composites at $t = 12$ h of (a) PV (PVU; shading) and potential temperature (black contours every 3 K) at 850 hPa and wind speed at 250 hPa (green contours for 50 and 60 m s^{-1}); (b) 6-h accumulated precipitation (mm; shading) and upper-level induced QG vertical motion at 700 hPa (black contours every 10^{-3} m s^{-1} , dashed for negative values); (c) potential temperature anomaly at 900 hPa (K; shading) and PV anomaly at 850 hPa (black contours from 0.5 PVU every 0.1 PVU), both obtained by subtracting the December–February climatological mean from the field at each location, frequency of WCB trajectories with $900 < p < 700$ hPa and $\text{PV} > 1$ PVU (green contours for 10%, 50%, and 65%) and their mean PV value (blue contours for 1.3, 1.4, and 1.5 PVU; only shown for areas where the green contour is $> 10\%$); and (d) brightness temperature of the false-color infrared satellite imagery ($^{\circ}\text{C}$; shading) derived from the GridSat-B1 data, overlaid by the 10% contours of low-level (black solid) and upper-level (black dashed) WCB trajectory frequencies. The gray cross marks the position of the cyclone center. Please note that (c) is enlarged to the $10^{\circ} \times 10^{\circ}$ box around the cyclone center.

WCB ascent (Fig. 7d), illustrating the key role of the WCB for the formation of the intense vortex. The zoom on the region around the cyclone center in Fig. 8c emphasizes the coincidence of the low-level PV anomaly (maximum of more than 1.2 PVU; see black contours) and the WCB-related PV production (maximum HLPV

WCB trajectory frequency of 65%, with a mean PV value of 1.5 PVU; see green and blue contours) with the cyclone center. Figure 8c also reveals that the near-surface potential temperature anomaly is positive in the cyclone center, but the strongest anomaly (more than 7 K) is located to the southeast of the cyclone center. This

indicates that, applying a conceptual piecewise PV inversion, both the lower-boundary warm anomaly and the low-level PV anomaly contribute to the near-surface cyclonic circulation, but the circulation induced by the low-level PV anomaly is stronger, as it is perfectly aligned with the overall cyclonic circulation. As shown in the supplemental material, this conclusion is supported by the fact that during cyclone intensification, low-level PV increases more strongly than the low-level potential temperature anomaly.

The low-level vortex is located at the leading edge of a pronounced cyclonically breaking upper-level wave (Figs. 7c,d) below the left exit region of an intense jet (Fig. 8a). Consistently, the upper-level induced QG vertical motion at 700 hPa reveals a very strong upper-level forcing for ascent in the region of the developing cyclone and the ascending WCB (Fig. 8b). The intense QG forcing is essential for the ascent of the WCB trajectories, while the low-PV air in the WCB outflow (see blue lines in Figs. 7c,d) most likely enhances the upper-level forcing by amplifying the downstream ridge. The existence and interplay of a strong upper-level forcing and an intense, diabatically produced low-level PV anomaly explains the explosive intensification of the C1 cyclones.

The occurrence of strong WCB-related cloud formation in the C1 cyclones is confirmed by satellite imagery. Figure 8d shows an infrared satellite composite of the C1 cyclones at $t = 12$ h. The satellite data are retrieved from the GridSat-B1 data (Knapp et al. 2011), which is available from January 1980 onward. From the 500 C1 cyclones all except the 13 cases occurring in 1979 are included in the composite. The composite reveals the presence of a cyclonically curved cloud band with a gradual decrease of brightness temperature, from warm, low-to-midlevel clouds south and southeast of the cyclone center, to cold, high-reaching clouds east and northeast. The cloud band has the characteristic comma shape often found in developing cyclones, with the comma head located above the storm center and at the warm front and the tail along the cold front. It exhibits a strong agreement with the comma-shaped precipitation signal (Fig. 8b) and the spatial distribution of the ascending WCB and, thus, provides observational evidence for the existence of a WCB cloud band associated with significant diabatic processes.

3) STRUCTURE AT $t = 24$ h AND SUBSEQUENT DEVELOPMENT

During the following 12 h, the composite cyclone further intensifies to a minimum SLP of about 972 hPa, accompanied by continuous low-level diabatic PV production, and the LC2-type rolling up of the upper-level

wave (Figs. 7e,f). A low-level vortex with PV values of more than 1.5 PVU extends up to about 550 hPa, with maximum values of more than 2 PVU near 850 hPa (Fig. 7f). The vortex is connected to a narrow filament of stratospheric air penetrating deep into the troposphere, and together they form an intense PV tower. It is at this time that the composite cyclone reaches its strongest intensity in terms of low-level relative vorticity (see supplemental material).

While the WCB trajectory frequency at low levels has slightly decreased compared to previous time steps, the WCB outflow has strengthened (Figs. 7e,f). The structure of the composite WCB outflow suggests that the WCB is associated with a cyclonic and an anticyclonic branch, as noted earlier for the case study (cf. Fig. 3e). In the west–east vertical section the cyclonically turning branch of the WCB outflow is visible between about 300 and 500 hPa slightly to the west of the low-level vortex, in a region where the upper-level trough is less pronounced and maybe eroded by diabatic processes (Fig. 7f). The anticyclonic branch east and northeast of the cyclone center reaches higher levels, consistent with the findings of Martínez-Alvarado et al. (2014), and contributes significantly to the amplification of the downstream ridge. As verified by choosing a larger compositing region (not shown), the ridge amplification in turn promotes Rossby wave breaking and the formation of a pronounced trough about 2500 km downstream of the composite cyclone. This suggests that the WCB associated with the C1 cyclones also plays a fundamental role for the wave dynamics at the tropopause level by strongly influencing the downstream flow evolution.

The strongest intensity in terms of SLP is reached 12 h later. By then the WCB-related PV production has strongly decreased, and the composite storm has become located below a deep tropopause (not shown).

Note that the structure and evolution of the composite cyclone exhibits strong similarities to the case study discussed above: pronounced WCB-related diabatic PV production in the cyclone center in a region of strong baroclinicity, a cyclonically breaking upper-level wave, and the formation of a PV tower toward the end of the intensification. An examination of the individual cases revealed that the composite well represents the main characteristics of C1 cyclones. For instance, two prominent C1 cyclones that have been investigated in detail, the Presidents' Day snowstorm and the windstorm Xynthia, both bear strong resemblance to the mean C1 composite (not shown). However, of course there is some case-to-case variability, with the importance of individual dynamical and physical processes varying among the cyclones. The most striking common feature

among the majority of the C1 cyclones is the concentration of high WCB-related PV production in the cyclone center during the explosive intensification. This implies that C1 cyclones are primarily associated with WCBs of type W2, which ascend close to the cyclone center.

5. Category C2—Weakly intensifying cyclones with strong WCBs

a. Spatial structure of the C2a cyclones

The composites of the 29 C2a cyclones at $t = 12$ h are shown in Fig. 9. The composite cyclone is associated with a weak small-scale SLP signal (Fig. 9a). It is located south of a strong baroclinic zone in a region of warm air (Fig. 9c) and a pronounced positive surface θ anomaly (Fig. 9e; note also the absence of a cold anomaly). An intense WCB is present (Figs. 9a,e), with a maximum WCB trajectory frequency at low levels of 80% slightly east of the cyclone center, a maximum HLPV trajectory frequency of 50% directly in the cyclone center, and a WCB outflow maximum of 60% to the northeast (similar to C1; cf. Figs. 7c and 8c). The area of strong PV production by HLPV WCB trajectories in and around the cyclone center (Figs. 9a,e) coincides with an intense low-level vortex with high PV values of more than 1 PVU (Figs. 9b,c) and intense precipitation (Fig. 9d). The low-level vortex extends up to about 700 hPa within a region of strong WCB ascent (Fig. 9b). It is not as strong as in C1 (cf. Fig. 7d) but still constitutes a pronounced positive PV anomaly (Fig. 9e). Farther to the east, the majority of the WCB trajectories is still located in the boundary layer and will ascend along the baroclinic zone during the subsequent hours (Fig. 9b).

In contrast to C1, the composite cyclone is located far south of a weak upper-level wave on 315 K (Fig. 9a), and above the shallow low-level vortex the tropopause is high and relatively undisturbed (Fig. 9b). Consistently, the upper-level forcing for ascent is very weak near the cyclone center (Fig. 9d), with about 3-times-lower values than in C1.

In the satellite composite of the 29 C2a cyclones, strong cloud formation is evident in the region of the ascending WCB (Fig. 9f), corroborating the importance of WCB-related cloud diabatic processes for these cyclones. As for C1, the brightness temperature gradually decreases from the cyclone's warm sector toward the area of strong WCB ascent east and northeast of the cyclone center. However, in C1 the cloud band is more cyclonically curved, and the clouds reach higher levels (cf. Fig. 8d). The lower cloud-top

height in C2a is in agreement with the lower height of the WCB outflow seen in Fig. 9b and likely a consequence of the weak upper-level forcing for ascent.

Over the subsequent hours the situation does not change substantially, and the composite cyclone intensifies only weakly (see supplemental Figs. S1, S2, and S4). Continuous WCB-related diabatic PV production further strengthens the low-level vortex, but the small-scale cyclone remains in a region of weak upper-level forcing for ascent below a relatively undisturbed tropopause (in contrast to C1, where an intense PV tower develops). The absence of a significant upper-level forcing during the entire development of C2a cyclones is the main difference from C1 and explains why they are not intensifying despite the strong WCB-related PV production.

The characteristics of the C2a cyclones bear strong resemblance to DRWs: a pronounced low-level positive PV anomaly in the center of a small-scale cyclone, its rapid eastward propagation along an intense baroclinic zone through continuous PV generation by ascending air to the east of the present low-level vortex, and the absence of a significant upper-level forcing. Indeed, among the C2a cyclones are four cases that have also been found in the group of nondeepening DRWs in the 10-yr climatology of Boettcher and Wernli (2013).⁴ The C2a cyclones therefore represent a subclass of nondeepening DRWs where the low-level vortex is generated by a strong WCB.

b. Spatial structure of the C2b cyclones

Figure 10 shows the composites of the 72 cyclones in the second subcategory of C2 at $t = 12$ h. A pronounced, meridionally elongated upper-level wave is located upstream of the composite cyclone (Fig. 10a), which, however, is characterized by a weak PV gradient in comparison to C1 (cf. Fig. 7a). Also the baroclinicity, the surface warm anomaly, and the upper-level flow in the region of the developing cyclone, as well as the upper-level induced QG vertical motion at 700 hPa, are weaker than in C1 (Figs. 10c–e).

Overall the C2b cyclones have almost the same WCB trajectory mass as C1, and a larger mass than C2a (see supplemental Fig. S1), but in C2b they are distributed over a larger area. In the cyclone center the WCB trajectory frequency is lower than in both C1 and C2a (but still has a maximum), and also the HLPV trajectory frequency is lower, whereas more trajectories are

⁴ However, the comparison is not straightforward, as Boettcher and Wernli (2013) used ECMWF operational analyses with a different model setup and higher resolution than ERA-Interim.

FIG. 9. C2a composites at $t = 12$ h. Fields as in (a) Fig. 7c, (b) Fig. 7d, and (c)–(f) Fig. 8.

FIG. 10. As in Fig. 9, but for C2b.

located relatively far away, particularly far south of the cyclone center (Fig. 10a). The WCB outflow is shifted north and slightly eastward with respect to the low-level trajectory positions.

In the cross section, a relatively broad upper-level trough is evident west of a low-level vortex with PV values of more than 1 PVU in the cyclone center (Fig. 10b). The latter indicates that also the C2b cyclones are associated with intense diabatic PV production. However, the vortex is weaker than in C1, with a positive PV anomaly of only 0.7 PVU at 850 hPa (Fig. 10e) compared to 1.2 PVU in C1, and also the vertical extent of the vortex is smaller (cf. Figs. 7d and 8c). This is consistent with the lower relative WCB trajectory density near the cyclone center and the smaller number of HLPV WCB trajectories (Figs. 10a,b,e). Interestingly, nonetheless the density of the upper-level trajectories above the low-level vortex is of about the same magnitude as in C1. The low-PV air in the WCB outflow directly above the cyclone center potentially impedes the stratospheric air to descend and connect with the low-level vortex.

Comparison of the spatial distribution of the WCB trajectories at low levels with the precipitation signal in Fig. 10d shows a strong correspondence between the two quantities: the maximum precipitation also occurs near the cyclone center, but it is weaker than in C1 and C2a, while an elongated band of precipitation connects the cyclone center with the border of the compositing region far to the southwest. It indicates cold-frontal precipitation, and the correspondence with the WCB trajectories implies that strong WCB ascent—and associated precipitation—occurs along the cold front in type W1 flows.

Note that the composites are not rotated with respect to the orientation of the cold front, and because of the variability in the cold frontal position among the individual cyclones its features are smeared out by the compositing. Therefore, the cold front appears relatively weak in the signal of θ (Fig. 10c), precipitation (Fig. 10d), and WCB trajectory frequency (Fig. 10a), but it is remarkable that nevertheless such a clear signal is evident, in contrast to C1 and particularly C2a. An examination of the individual C2b cyclones revealed that in most cases the WCB ascent and associated PV production concentrated along a pronounced cold front far from the cyclone center (not shown). The cold front was typically associated with a narrow stratospheric PV streamer or a cutoff. The orientation of the PV streamer varied between LC1 type, neutral, and LC2 type, and this variability can explain the relatively broad upper-level disturbance in the composite (Fig. 10a).

In the satellite composite (Fig. 10f; containing the 69 cyclones occurring from 1980 onward) the cold frontal

cloud band is indicated by increasing brightness temperatures from the southern edge of the compositing region toward the cyclone center. It nicely correlates with the spatial distribution of the ascending WCB trajectories. The highest-reaching clouds are located in the region of strongest WCB ascent slightly northeast of the cyclone center. Their cloud top is lower than in C1, consistent with the weaker upper-level forcing for ascent.

During the following hours, the described fields remain very similar, and the composite cyclone does not intensify considerably (see supplemental Figs. S1, S2, and S5). The westward tilt between the cyclone center and the upper-level trough remains almost constant, and no PV tower develops, contrary to C1.

In summary, the absence of intensification in the C2b category despite the strong WCB can be explained by several factors: (i) In contrast to C1, much of the WCB ascent and associated PV production occurs in many cases at the cold front far south of the cyclone center, in WCBs of type W1. The fraction ascending near the cyclone center is much lower and associated with a weaker diabatically produced vortex. This indicates that the location of the WCB-related PV production relative to the cyclone center plays an important role for its impact on cyclone development. (ii) The upper-level disturbance has a weaker PV gradient than in C1, and accordingly the upper-level forcing for ascent is weaker. (iii) Directly above the low-level vortex, the density of the upper-level WCB trajectories is higher than in C1, particularly toward the end of intensification, which potentially additionally impedes the formation of a PV tower.

6. Category C3—Explosively intensifying cyclones with weak WCBs

The mean spatial structure of the 154 C3 cyclones at $t = 12$ h is shown in Fig. 11. The SLP field reveals a small-scale low pressure system in the center of the composite and another system to the northeast (Fig. 11a). In agreement with the high latitude of most C3 cyclones (Fig. 2), the tropopause is low (Fig. 11a) and it is considerably colder than in the other categories (Fig. 11c). The positive surface θ anomaly in the warm sector is relatively weak, while the cold sector is associated with a pronounced negative anomaly (Fig. 11e). The composite cyclone is located at the left jet exit in a region of relatively strong low-level baroclinicity (Fig. 11c). Both fields are pronounced, but weaker than in C1 (cf. Fig. 8a). Upstream of the cyclone center, a small disturbance is present on the 315-K isentrope that is associated with a strong PV gradient (Fig. 11a). In the

FIG. 11. As in Fig. 9, but for C3.

cross section a deep trough is evident east of the cyclone (Fig. 11b). The upper-level forced QG ascent at 700 hPa is weaker than in C1 and of about the same magnitude as in C2b (Fig. 11d).

As expected, the WCB is weak, with a maximum WCB trajectory frequency of 20% at low levels near the cyclone center, a 10% frequency of HLPV WCB trajectories in the same region, and a 5% frequency in the WCB outflow slightly to the north (Figs. 11a,e). The precipitation signal is much weaker than in the other categories (Fig. 11d), and also the comma-shaped cloud in the satellite composite is at lower height (Fig. 11f; containing 151 C3 cyclones). Nevertheless, above the cyclone center a low-level vortex of a similar strength as in C2a and C2b indicates that also in C3 diabatic processes play a role (Figs. 11b,c). The small number of WCB trajectories suggests that the low-level PV production is to a large extent related to diabatic processes that are independent of WCBs. Note, however, that the PV anomaly of the low-level vortex amounts to only 0.5 PVU (Fig. 11e), which is less than in the other categories, particularly C1 and C2a.

Over the next hours, the composite cyclone strongly intensifies and grows in size at the expense of the cyclone to the northeast (see supplemental Fig. S6). The pattern suggests a merging of the two systems, which would explain the rapid SLP decrease. This indeed occurs in some C3 cyclones, but often they remain isolated and simply develop in the vicinity of another, decaying system, which appears as a merging signal in the composite. Toward the end of the intensification, a weak PV tower develops through a further lowering of the tropopause and ongoing diabatic processes.

For the C3 cyclones at high latitudes ($>60^\circ\text{N}$ in Fig. 2), it is noted that they can be located in regions where the troposphere is lower than the pressure threshold of 600 hPa used to identify WCBs. This raises the question whether these cyclones are nevertheless connected to strong diabatic processes in airstreams ascending less than 600 hPa. However, case studies showed that the high-latitude C3 cyclones are associated with very low values of low-level specific humidity, weak precipitation, and accordingly weak diabatic processes. This confirms that the weak WCB strength in these cyclones is not simply due to the 600-hPa WCB ascent criterion used in this study.

To summarize, C3 cyclones are associated with a relatively pronounced baroclinicity, jet and upper-level forcing, and some low-level diabatic PV production. However, all these fields are weaker than in C1, which can in part be attributed to the smaller mean Bergeron value of C3, as discussed in section 3. Furthermore, an

examination of the individual C3 cyclones revealed that the category is rather inhomogeneous, with various types of explosive intensification.

7. Summary and conclusions

The present study comprises a first systematic analysis of the statistical relationship between cyclone intensification and WCB strength. The analysis is based on WCB and cyclone climatologies compiled from the ERA-Interim dataset for the NH winter during 1979–2014. For the ~ 5000 cyclones in the dataset, it has been investigated whether there is a connection between cyclone intensification and the strength of the associated WCB. Cyclone intensification has been quantified as the latitude-adjusted maximum central SLP deepening over 24 h along a track, as introduced by Sanders and Gyakum (1980). The WCB strength has been measured by the number and mass of the WCB trajectories associated with the cyclone at low levels (pressure > 500 hPa) during the 24 h of strongest SLP deepening. Based on the values for cyclone intensification and WCB strength, three cyclone categories have been distinguished: explosively intensifying cyclones with a strong WCB (C1), weak cyclones with a strong WCB (C2), and explosively intensifying cyclones with a weak WCB (C3). Their structure and evolution has been analyzed using composites. The questions posed in the introduction can be answered as follows:

- 1) The statistical analysis of the relationship between cyclone intensification and WCB strength yields a Spearman correlation coefficient of 0.68. This implies that there is statistically a clear signal that stronger intensifying cyclones have stronger WCBs, which contribute to the diabatic production of cyclonic low-level PV. This result is in harmony with the results of many case studies (see introduction) and with the climatological analysis of Campa and Wernli (2012) that low-level PV production due to latent heating in clouds is typically enhanced in explosive cyclones and contributes to the formation of the lower part of the cyclones' PV towers. An important new result of this study is that this low-level PV production occurs essentially in WCBs. However, the majority of the cyclones is associated with a weak deepening and few or no WCB trajectories, while the strongly deepening cyclones with an intense WCB occur less frequently.
- 2) Explosively intensifying cyclones with a strong WCB (C1) are typically characterized by intense PV production in the cyclone center related to a WCB of type W2, the interaction of the diabatically produced low-level vortex with a cyclonically breaking upper-level wave

and a surface warm anomaly, and the formation of an upright PV tower toward the end of strongest intensification. The diabatically generated PV anomaly within the ascending WCB significantly enhances the low-level circulation and thereby contributes to the explosive intensification.

- 3)(i) Cyclones with a similar WCB strength as in C1 but a weak deepening rate (C2) are divided into two subcategories:

C2a: DRW-like cyclones associated with strong low-level PV production by WCBs in the cyclone center in a baroclinic region, but a weak upper-level forcing for ascent, which prevents the formation of an intense cyclone. The existence of WCB airstreams in DRWs despite the absence of significant upper-level forcing is particularly interesting, as it implies that in these cases the strong ascent must be almost entirely driven by the slantwise circulation over the baroclinic region induced by the low-level vortex.

C2b: Cyclones with substantial PV production along the pronounced cold front far away from the cyclone center in WCBs of type W1. This results in a weaker low-level PV anomaly in the cyclone center in comparison to C1, and in addition, the cyclones are associated with a weaker upper-level forcing for ascent. For these cyclones the diabatic PV production in the WCB and the dry dynamical ingredients are not in phase, which inhibits rapid intensification. Further investigations are needed to better understand this behavior, which however occurs much less frequently than the synergetic phasing of the anomalies (500 cyclones in category C1 vs only 72 in category C2b).

- 3)(ii) The category of explosively intensifying cyclones with a weak WCB (C3) is inhomogeneous and can approximately be classified as follows: (i) particularly strong upper-level forcing for ascent (e.g., at the left jet exit); (ii) important role of WCB-related PV production, although the WCB is associated with few trajectories; (iii) important role of diabatic PV production not related to WCBs; and (iv) merging of cyclones leading to a rapid SLP decrease. Often, the cyclones also reveal a combination of these characteristics.

The findings for the different categories are fairly robust to the measure used to quantify cyclone intensification, as shown in the supplemental material. In addition, very similar results are obtained when using as a measure of WCB strength the number of WCB trajectories normalized by the size of the cyclones (not shown). Two

aspects of the previously summarized key findings deserve a brief further discussion: (i) the relative importance of W1 and W2 WCBs for cyclone intensification and (ii) the link to moist baroclinic instability studies.

In many explosively intensifying cyclones (category C1) there is a remarkable concentration of the WCB-related PV production in the cyclone center. This shows that these cyclones are predominantly associated with a WCB of type W2, which ascends close to the cyclone center above the bent-back front (e.g., [Browning and Roberts 1994](#); [Wernli 1997](#)). The results of this study imply that the ascent in the cyclone center, which leads to the type W2 WCB, is strongly determined by the QG upper-level forcing for vertical motion ([Fig. 8b](#)). The WCB associated with the weakly intensifying C2b cyclones, on the other hand, fits well with the description of type W1, particularly the part of the WCB ascending far away from the cyclone center. This demonstrates that WCBs of type W1 and W2 are dynamically distinct flows that have strongly differing effects on cyclone intensification. W2 WCBs significantly enhance cyclone intensification through intense low-level PV production close to the cyclone center, in contrast to W1 WCBs, where the low-level PV production occurs too far away from the cyclone center. The position of the WCB ascent relative to the cyclone center, and the processes determining this position, therefore play a crucial role for the evolution of the cyclone.

Establishing a link between our results and idealized baroclinic instability studies (e.g., [De Vries et al. 2010](#); [Cohen and Boos 2016](#)) is not straightforward mainly because we focus on the period of maximum intensification when finite-amplitude anomalies interact nonlinearly. For C1 cyclones, the strongest growth during the 24 h of maximum deepening occurs for the low-level PV anomaly ([Figs. 7](#) and [S1](#)). The upper-level PV anomaly also shows considerable intensification ([Fig. 7](#)) while the surface potential temperature anomaly intensifies much less ([Fig. S2](#)). This indicates that for C1 cyclones the favorable interaction between upper-level (dry) PV and low-level (diabatically generated) PV is essential, which was referred to as a “tropopause intrusion” by [Cohen and Boos \(2016\)](#) and as “type-C baroclinic growth” by [De Vries et al. \(2010\)](#). The latter terminology would be consistent with [Plant et al. \(2003\)](#) and [Ahmadi-Givi et al. \(2004\)](#), who also emphasized the interaction of prominent upper- and low-level PV anomalies for type C cyclones while surface thermal anomalies remain weak. For C2a cyclones, the most important anomalies are clearly those of low-level PV and surface potential temperature, which we consider as typical for diabatic Rossby waves (e.g., [Boettcher and Wernli 2011](#)). This however differs from the diabatic

Rossby wave definition in the more idealized studies cited above. For C2b and C3, a one-to-one attribution with any of the idealized categories is even more difficult.

Finally, it is noted as a caveat that this study is based on relatively coarse-resolution reanalysis data, and that with the compositing technique, we focused on robust PV structures on the subsynoptic scale (~ 500 km). It is however known that cyclones also reveal important mesoscale PV structures (e.g., Thorpe and Clough 1991; Neiman et al. 1993). Investigation of the formation of these PV anomalies and their relationship to the WCB and its associated microphysical processes requires the use of a high-resolution mesoscale model. This route of research has so far only been applied for selected case studies (e.g., Joos and Wernli 2012). In addition, the approach used in this climatological study is potentially useful for quantitatively assessing cyclone dynamics in climate models. Catto et al. (2010) analyzed whether climate models can capture the overall structure of extratropical cyclones; the framework developed here could serve to analyze the interaction of latent heating and cyclone dynamics in these models in greater detail.

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