A New Perspective toward Cataloging Northern Hemisphere Rossby Wave Breaking on the Dynamic Tropopause

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ABSTRACT

Rossby wave breaking (RWB) events are a common feature on the dynamic tropopause and act to modulate synoptic-scale jet dynamics. These events are characterized on the dynamic tropopause by an irreversible overturning of isentropes and are coupled to troposphere-deep vertical motions and geopotential height anomalies. Prior climatologies have focused on the poleward streamer, the equatorward streamer, or the reversal in potential temperature gradient between the streamers, resulting in differences in the frequencies of RWB. Here, a new approach toward cataloging these events that captures both streamers is applied to the National Centers for Environmental Prediction Reanalysis-2 dataset for 1979–2011. Anticyclonic RWB (AWB) events are found to be nearly twice as frequent as cyclonic RWB (CWB) events. Seasonal decompositions of the annual mean find AWB to be most common in summer (40% occurrence), which is likely due to the Asian monsoon, while CWB is most frequent in winter (22.5%) and is likely due to the equatorward shift in mean baroclinicity. Trends in RWB from 1980 to 2010 illustrate a westward shift in North Pacific AWB during winter and summer (up to 0.4% yr$^{-1}$), while CWB in the North Pacific increases in winter and spring (up to 0.2% yr$^{-1}$). These changes are hypothesized to be associated with localized changes in the two-way interaction between the jet and RWB. The interannual variability of AWB and CWB is also explored, and a notable modality to the frequency of RWB is found that may be attributable to known low-frequency modes of variability including the Arctic Oscillation, the North Atlantic Oscillation, and the Pacific–North American pattern.

1. Introduction

Diagnosing and interpreting the importance of Rossby wave breaking (RWB) events for midlatitude high- and low-frequency variability has been an area of considerable research (e.g., Hoskins et al. 1985; Nakamura et al. 1997; Martius et al. 2006; Michel and Rivière 2011). RWB has been examined in reanalysis datasets (e.g., Postel and Hitchman 1999; Scott and Cammas 2002; Benedict et al. 2004; Abatzoglou and Magnusdottir 2006; Wernli and Sprenger 2007; Liu et al. 2014), future climate simulations (e.g., Rivière 2011; Barnes and Hartmann 2012), and idealized model simulations (e.g., Thorncroft et al. 1993; Nakamura and Plumb 1994; Peters and Waugh 1996). Such studies have also utilized a variety of methodologies for identifying and cataloging RWB (e.g., Postel and Hitchman 1999; Waugh and Polvani 2000; Martius et al. 2007; Hitchman and Huesmann 2007; Barnes and Hartmann 2012; Liu et al. 2014). In general, there is agreement on the climatological locations of RWB but discrepancies in the frequency, with differences of up to double the frequency from one study to another (e.g., Martius and Rivière 2016). Such differences can influence the interpretation of trends in RWB and their relative role in low-frequency variability. Therefore, our goal is to propose a new approach toward defining and cataloging RWB events that consolidates the results of these previous studies.

Synoptic-scale Rossby waves break when they become highly amplified and their phase speed matches that of the background flow (e.g., Polvani and Plumb 1992; Nakamura and Plumb 1994; Swanson et al. 1997), at which point substantial meridional transport of air masses occurs in the stratosphere, troposphere, or both (e.g., McIntyre and Palmer 1983; Polvani and Plumb 1992; Hitchman and Huesmann 2007; Liu and Barnes 2015). Breaking Rossby waves are characterized by a poleward intrusion of low potential vorticity (or high...
potential temperature) air and an equatorward intrusion of high potential vorticity (or low potential temperature) air that is dictated by the shear environment associated with the incipient Rossby wave (Thorncroft et al. 1993). These midlatitude flow reconfigurations, which result in the displacement of filamentary regions of high potential vorticity air equatorward and low potential vorticity air poleward, are frequently referred to as potential vorticity streamers (e.g., Shapiro 1980; Keyser and Shapiro 1986; Appenzeller and Davies 1992; Wernli and Sprenger 2007).

When exposed to a cyclonically sheared environment, potential vorticity streamers can undergo cyclonic (LC2, P1) wave breaking (CWB) wherein a low potential vorticity streamer is displaced poleward and eastward of a high potential vorticity streamer. In an anticyclonically sheared environment, anticyclonic (LC1, P2) wave breaking (AWB) of potential vorticity streamers results in a low potential vorticity streamer being displaced poleward and westward of a high potential vorticity streamer. Differences between the LC-type and P-type wave breaking classifications lie in that LC-type events are considered to initiate from an equatorward-intruding trough (Thorncroft et al. 1993, their Fig. 12), while P-type events are considered to initiate from a poleward-intruding ridge (Peters and Waugh 1996, their Fig. 5). Both instances result in a localized reversal of the typical pole-to-equator potential vorticity gradient (e.g., Hoskins et al. 1985; Appenzeller and Davies 1992; Thorncroft et al. 1993; Peters and Waugh 2003; Martius et al. 2007; Gabriel and Peters 2008; Ndarana and Waugh 2011). These sheared environments are generally found on the equatorward and poleward flanks of jet streams, and as such, CWB is most frequent on the poleward flanks of the climatological jet stream, while AWB is found on the equatorward flanks (e.g., Abatzoglou and Magnusdottir 2006; Martius et al. 2007; Strong and Magnusdottir 2008b; Barnes and Hartmann 2012). Here, we are less interested in the initializing feature (trough or ridge) than the more universally consistent sense of background shear (cyclonic or anticyclonic) and as such will simply classify events as either CWB or AWB.

The identification and classification of RWB has been approached using several different techniques. Baldwin and Holton (1988) pioneered the first automated RWB identification technique based upon the condition of localized reversals of the potential vorticity gradient. This technique was used as the basis for several studies of tropospheric and stratospheric RWB (e.g., Postel and Hitchman 1999; Waugh and Polvani 2000; Hitchman and Huesmann 2007). More recently, contour searching algorithms have been applied to identify overturning contours of potential vorticity or potential temperature (e.g., Abatzoglou and Magnusdottir 2006; Martius et al. 2007; Barnes and Hartmann 2012; Liu et al. 2014). Further differences in methodology are found with the choice of coordinate system, which includes the analysis of potential vorticity on isentropic surfaces (e.g., Postel and Hitchman 1999; Peters and Waugh 2003; Abatzoglou and Magnusdottir 2006; Wernli and Sprenger 2007; Hitchman and Huesmann 2007), isentropes on potential vorticity surfaces (e.g., Benedict et al. 2004; Liu et al. 2014), or absolute vorticity on pressure surfaces (e.g., Rivière 2009; Rivière et al. 2010; Rivière 2011; Barnes and Hartmann 2012). The frequency of RWB detection can be dependent on the surface analyzed, where CWB is more frequent on colder isentropic levels (e.g., Abatzoglou and Magnusdottir 2006; Martius et al. 2007), while AWB is more frequent on warmer isentropic levels (e.g., Abatzoglou and Magnusdottir 2006). Hitchman and Huesmann (2007) demonstrated the importance of identifying potential vorticity reversals on a series of isentropic surfaces (320–2000 K) and also show equatorward shift in RWB toward warmer isentropic surfaces near the tropopause.

Climatologies of tropospheric RWB in the Northern Hemisphere have identified several regions of enhanced breaking activity. The majority of studies have found these regions to be clustered over the North Pacific and North Atlantic oceanic basins (e.g., Postel and Hitchman 2001; Scott and Cammas 2002; Abatzoglou and Magnusdottir 2006), though some have also found extensions of these regions eastward into portions of Europe and Asia (Martius et al. 2007; Strong and Magnusdottir 2008b; Barnes and Hartmann 2012). These clusterings are predominantly found in association with the enhanced horizontal shear associated with subtropical and mid-latitude jets (e.g., Thorncroft et al. 1993). Complex two-way interactions between the jet stream and RWB can also act to perturb the location of the jet and modulate the frequency of Rossby wave breaking. Rivière (2009) showed that the latitudinal position of the jet can robustly impact the frequency of RWB, wherein a poleward deflection of the jet favors more AWB and an equatorward deflection favors a higher frequency of CWB. However, momentum transfer associated with AWB has been shown to deposit momentum poleward of the breaking wave, while CWB deposits momentum equatorward, resulting in a poleward and equatorward displacement of the jet, respectively (e.g., Thorncroft et al. 1993; Orlanski 2003; Rivière 2009). This two-way process adds complexity when attempting to diagnose changes in RWB frequencies that are collocated with changes in the jet position.

Generally, localized maxima in RWB frequencies translate from eastern portions of the North Atlantic and North Pacific basins in winter to central and western
portions of the basins in summer (Scott and Cammas 2002; Abatzoglou and Magnusdottir 2006). Winter AWB has been observed to be more common in the North Atlantic basin than the North Pacific basin, while the opposite is true of CWB (Martius et al. 2007; Strong and Magnusdottir 2008b). Differences between the basins for winter RWB frequency can be a result of the orientation of the seasonal-mean jet structure, which favors a split jet in the North Atlantic and a single jet in the North Pacific (Abatzoglou and Magnusdottir 2006), or a more equatorward jet in the North Pacific, compared to the North Atlantic (Akahori and Yoden 1997; Rivière 2009). Changes in winter RWB also act to modulate the North Atlantic Oscillation (NAO) in that momentum fluxes associated with North Pacific AWB events can act to instigate North Atlantic AWB events, which project onto the positive phase of the NAO (e.g., Benedict et al. 2004; Abatzoglou and Magnusdottir 2006; Strong and Magnusdottir 2008a; Woolings et al. 2008). Such changes can also be seen in the northern annular mode (NAM), which can result in poleward or equatorward shifts in the jet favoring AWB and CWB, respectively (e.g., Drouard et al. 2015). Furthermore, the positive phase of the winter Pacific–North American pattern (PNA) have been found to be associated with anomalously high AWB and low CWB in the North Pacific, while the opposite is true for the negative phase (Franzke et al. 2011).

Anticyclonic wave breaking is found to be most frequent during the summer (Postel and Hitchman 2001; Scott and Cammas 2002; Abatzoglou and Magnusdottir 2006) and is maximized in the North Pacific basin due to the strength of the Asian monsoon (Postel and Hitchman 1999; Abatzoglou and Magnusdottir 2006). Autumn (SON) AWB is found to be more common than spring AWB in both the North Pacific and North Atlantic basins, which is thought to be due to a poleward shift of the autumn critical latitude (where waves become stationary relative to the background flow; Scott and Cammas 2002) and the SON jet stream being displaced upward of 10° north of that in the spring (Abatzoglou and Magnusdottir 2006). The El Niño–Southern Oscillation (ENSO) has also been shown to play an important role in modulating RWB, during which the warm phase is found to be associated with enhanced CWB (e.g., Shapiro et al. 2001; Abatzoglou and Magnusdottir 2006; Martius et al. 2007; Liu et al. 2014).

Despite the agreement of several studies on the climatological regions of RWB, there are clear biases that can occur based on the selection of technique. Studies that identify RWB events using potential vorticity gradients on isentropic surfaces are sensitive to the selection of isentropes for analysis (e.g., Postel and Hitchman 1999; Scott and Cammas 2002; Abatzoglou and Magnusdottir 2006; Martius et al. 2007). Further differences exist as to the identification of an RWB, wherein some studies identify the equatorward potential vorticity streamer (e.g., Wernli and Sprenger 2007; Martius et al. 2007; Isotta et al. 2008), the poleward potential vorticity streamer (e.g., Strong and Magnusdottir 2008b; Liu et al. 2014), and the region of potential vorticity gradient reversal (e.g., Postel and Hitchman 1999; Scott and Cammas 2002). Such differences can result in the identification and cataloging of different regions of an RWB event as well as different points in the event life cycle, resulting in variations in the frequency of detection.

Here, we present a different approach that can be applied toward the automated identification and counting of RWB events in the Northern Hemisphere. Events are identified by searching along numerous isentropes on the dynamic tropopause [DT; defined as the 2.0 potential vorticity unit (PVU; 1 PVU = 1.0 × 10⁻⁶ m² s⁻¹ K kg⁻¹] surface], allowing for the identification of tropopause RWB events throughout the extratropical Northern Hemisphere. The DT crosses from higher to lower isentropes with latitude (e.g., Martius et al. 2007), which can capture some of the differences in RWB frequencies between different isentropic surfaces. By identifying both the high and low potential temperature streamers associated with an RWB event, we believe that our approach offers a more generalized approach toward identifying and cataloging the regions of RWB. The methodology of identification and counting will be presented in section 2; a climatology of AWB and CWB will be presented in section 3; and 30-yr trends and the interannual variability in AWB and CWB will be presented in sections 4 and 5, respectively.

2. Methods

a. Data

Six-hourly data from the National Centers for Environmental Prediction (NCEP)–Department of Energy (DOE) Reanalysis-2 global reanalysis dataset (Kanamitsu et al. 2002) are used for this study for the period of 1 January 1979–31 December 2011. The horizontal resolution is 2.5° × 2.5° with 17 pressure levels ranging from 1000 to 10 hPa. To identify the DT, we first perform a linear interpolation to calculate potential temperature for the full depth of the reanalysis domain (1000–10 hPa). Any values of potential temperature that exceed 460 K are set as a missing value, as such isentropic surfaces are, on average, in the lower stratosphere (e.g., Hitchman and Huesmann 2007). This is done to avoid misrepresentations of the DT close to the equator, where potential vorticity approaches zero as the Coriolis
force approaches zero (under zonal flow with no background relative vorticity). This limits the focus of this study to extratropical tropopause RWB events, whereas previous studies have demonstrated the importance of studying a deeper layer of the stratosphere when examining troposphere–stratosphere interactions and stratospheric RWB (e.g., Baldwin and Holton 1988; Hitchman and Huesmann 2007). Next, we compute the potential vorticity in each isentropic layer and search for the first instance of a value of 2.0 PVU starting from the top of the atmosphere downward. The DT surface is defined in our study as the first identified 2.0-PVU surface.

Standardized monthly teleconnection indices are used to calculate seasonal-mean teleconnection indices for the 1980–2010 period that are regressed against seasonal-mean RWB frequencies. The standardized anomalies are calculated by the NOAA Climate Prediction Center (CPC) and are standardized using the 1981–2010 climatology. The data for this study include monthly indices for the NAO, the East Pacific pattern (EP), and the PNA and are calculated based on Barnston and Livezey (1987).

Seasonal-mean indices for the Arctic Oscillation (AO) were calculated by the CPC using the leading empirical orthogonal function (EOF) mode of the monthly mean 1000-hPa height anomalies from 20°S to 90°N. In addition, monthly values of the CPC Niño-3.4 index are utilized to calculate a seasonal-mean Niño-3.4 index.

b. Rossby wave breaking event identification

The identification of RWB events presented here is based in part upon the overturning contour identification technique developed by Barnes and Hartmann (2012). However, our method of identification searches for overturning isentropes on the DT, whereas they applied their technique to contours of absolute vorticity on pressure surfaces. The techniques are dynamically consistent, however, as low potential temperature streamers on the DT are generally equivalent to high potential vorticity streamers on isentropic surfaces (Hoskins et al. 1985; Hoskins 1997). The advantage of utilizing the DT is that it provides an ideal surface for analysis of jet stream–level processes as the amplitude of warm-core high and cold-core low pressure systems are maximized on this surface. These features shift with the DT across all seasons (Hoskins et al. 1985), allowing for a seasonally independent analysis along a single surface reducing computational expense. Hitchman and Huesmann (2007) showed cross-sectional analysis of potential vorticity reversal frequency and strength indicating that, on average, the DT broadly intersects the region of the strongest and most frequent reversals that are commonly associated with RWB along the tropopause.

An example of AWB over the western North Atlantic basin from 3 January 1979 is presented in Fig. 1 as a schematic guide for the diagnosis of RWB events. To begin, the method first seeks out continuous closed contours of potential temperature (5-K interval from 280 to 370 K) on the DT that encircle the hemisphere for each time step. Instances where the isentropes do not encircle the hemisphere are removed from the analysis, as well as closed contours that are separate from a contour that encircles the hemisphere. One such example in Fig. 1 is the 330-K closed contour near 27°N, 52°W, which would be removed from the analysis. However, the remainder of the 330-K surface that does encircle the hemisphere in this case would not be removed from the analysis. Such contours that do not encircle the hemisphere are discarded, as they may be associated with phenomena such as closed or cutoff high or low pressure regions (Hoskins et al. 1985; Wernli and Sprenger 2007).

Next, the method seeks any region of overturning contours, which are denoted by a single continuous isentrope crossing the same meridian at three different longitudes. This is best illustrated by examining the 50°W meridian in Fig. 1, where overturning is identified for the 335–355-K isentropes. After identifying the westernmost (starting) and easternmost (ending) overturning points on an isentrope (white circles in Fig. 1), the entire length of the contour between these two points is identified as an overturning contour (dashed contour in Fig. 1). To capture events that are approximately of synoptic scale, an overturning contour must meet the following criteria: (i) the contour must exceed 1500 km in total length, (ii) the region of overturning must exceed 5° longitude in horizontal extent, and (iii) the region of overturning cannot exceed 40° longitude in horizontal extent. In doing so, the method limits identified waves to typical synoptic scales in the midlatitudes (~2000 km). Any regions of overturning that do not meet these criteria are not identified as an overturning contour.

To classify an RWB event, the centroids of at least three overturning contours must fall within 15° great circle distance of each other, where the centroid for each overturning contour is defined as the mean latitude and longitude of all overturning points along a contour. In Fig. 1, the centroids of the 335–355-K contours (on the DT) meet this criteria and therefore are considered part of the same RWB event. The centroid of the RWB event is computed as the geographical center of all overturning contours classified as part of a breaking wave, and it is shown as the white star in Fig. 1. The type of RWB (anticyclonic or cyclonic) is determined following the technique of Barnes and Hartmann (2012) by comparing the respective latitudes of the starting and ending points along an overturning contour (white circles in...
If the starting (westernmost) point is poleward of the ending (easternmost) point, then the high potential temperature streamer is westward of the low potential temperature streamer, and the event is therefore typed AWB in the Northern Hemisphere. This is readily seen for the 335-K overturning contour in Fig. 1, for example, where the starting point is at 59°N (58°W) and the ending point is at 33°N (30°W). If the starting point is equatorward of the ending point, the event is typed CWB.

c. Generalized cataloging of regions of wave breaking

The identification and cataloging of RWB events has been approached differently by a variety of studies. Recent overturning contour identification methods have generally identified the RWB region as the breaking of high potential temperature (low potential vorticity) air protruding poleward (e.g., Strong and Magnusdottir 2008b; Liu et al. 2014), the breaking of low potential temperature (high potential vorticity) air protruding equatorward (e.g., Wernli and Sprenger 2007; Martius et al. 2007; Isotta et al. 2008), or the region of absolute vorticity reversal on isobaric or potential vorticity reversal isentropic surfaces (e.g., Hitchman and Huesmann 2007; Rivière 2009; Michel and Rivière 2011). However, both theory (e.g., Hoskins et al. 1985) and observations (e.g., Gabriel and Peters 2008; Strong and Magnusdottir 2008b) indicate that dynamically relevant processes occur across the entire wavelength of an RWB event. Therefore, we now introduce a generalized identification method that identifies both the equatorward- and poleward-protruding potential temperature anomalies (the full wave breaking wavelength) and more broadly captures the entire localized tropospheric response to RWB that may have only been partially captured by previous techniques.

To capture the full signature of an RWB event, our generalized method identifies the northernmost, easternmost, southernmost, and westernmost overturning points along the contours identified for an event (white diamonds in Fig. 1). The northernmost and easternmost points in the example presented in Fig. 1 are found along the 335-K isentropic contour, while the southernmost and westernmost points are found along the 355-K isentrope. These points are matched to the nearest grid point for the reanalysis grid (2.5°x2.5° grid spacing) and form the bounds that we define as our wave breaking region (denoted by the black box in Fig. 1). All points within the wave breaking region are identified as undergoing RWB and are utilized to compute the climatologies presented in section 3. Furthermore, the wave breaking region should capture the majority of the

Fig. 1. RWB event identification scheme applied to diagnose an AWB event from 3 Jan 1979. Gray shading denotes DT (2.0-PVU surface) isentropes in 5-K intervals, dashed lines represent overturning isentropes, and white circles denote the starting (westernmost) and ending (easternmost) points of each overturning isentrope (contour). The white star represents the centroid (geographical center) of all identified overturning isentropes for an event; the white diamonds represent the northern, eastern, southern, and westernmost bounds of the wave breaking event; and the black box is the identified wave breaking region rounded to the nearest reanalysis grid point.
localized quasigeostrophic forcing in response to the RWB event (e.g., Figs. 2, 3).

To illustrate the ability of our method to capture the effects of RWB through the depth of the troposphere, two events are presented: an AWB event from 1 December 1979 (Fig. 2) and a CWB event from 2 February 2010 (Fig. 3). Anticyclonic wave breaking is noted in Fig. 2a by a poleward streamer of high potential temperature air, a displaced streamer of low potential temperature air offset to the southeast along the date line, and a reversal of the climatological potential temperature gradient located at approximately 30°N from 175°E to 170°W. A region of 40 m s⁻¹ northeasterly winds divides the anticyclonic circulation in association with the high potential temperature region and the cyclonic circulation in association with the low potential temperature region (Fig. 2a). The signal of the event is also notable in the lower troposphere. A region of positive anomalies in 1000–500-hPa thickness (180 m; Fig. 2b) and sea level pressure (24 hPa; Fig. 2c) is found in association with the poleward-surging high potential temperature streamer on the northern flank of the wave breaking region. Negative anomalies in 1000–500-hPa thickness (−60 m; Fig. 2b) and sea level pressure (−8 hPa; Fig. 2c) are also noted in the region of the low potential temperature streamer in the southern flank of the wave breaking region.

The CWB event presented in Fig. 3a is characterized by a cyclonically sheared region of low potential temperature air extending off the coast of Atlantic Canada with a poleward streamer of anticyclonically sheared high potential temperature air displaced to the northeast near eastern Greenland. The reversal in potential temperature gradient is found near 60°N from 65° to 25°W and is characterized by 20–30 m s⁻¹ southeasterly winds (Fig. 3a). A 280-m positive anomaly in 1000–500-hPa thickness is found in the northern portion of the wave breaking region, which is coupled with a −240-m negative thickness anomaly in the southern portion of the wave breaking region (Fig. 3b). Furthermore, a negative sea level pressure anomaly is observed near 57°N, 48°W, and a positive anomaly near 65°N, 15°W (Fig. 3c). These anomalies are found downstream of regions of cyclonic and anticyclonic vorticity advection, respectively, that are associated with the potential vorticity streamers.

In both the AWB and CWB cases, it is clear that regions along the full wavelength of tropopause RWB are subject to troposphere-deep dynamical responses. It is evident that the wave breaking region (black box) in these cases captures the dominant surface response to AWB on the DT (a surface anticyclone) and CWB (a surface cyclone) that likely occurs in response to the enhanced vorticity advections associated with the
compressed half wavelength between the DT trough and ridge (Figs. 2, 3). However, these figures also illustrate that the wave breaking region broadly captures the weaker but still important surface cyclone found equatorward of the AWB (Fig. 2) and surface anticyclone found poleward of the CWB (Fig. 3). Furthermore, our events are similar to the tropospheric features of RWB found by Strong and Magnusdottir (2008b, their Fig. 3) in a composite of approximately 19,000 instantaneous winter AWB events and 12,000 CWB events.

3. Rossby wave breaking climatology

Rossby wave breaking events with centroids between 20° and 80°N are cataloged to create a climatology of AWB and CWB. The climatology is presented as the percentage of time during the reanalysis period for which a grid point is classified as being in an instantaneous wave breaking region. The annual-mean RWB frequency is shown in Fig. 4. Our results show that AWB is generally confined to the equatorward side of the climatological-mean jet stream and CWB to the poleward side, as has been shown in previous studies (e.g., Abatzoglou and Magnusdottir 2006; Martius et al. 2007; Strong and Magnusdottir 2008b; Barnes and Hartmann 2012). We also find that RWB frequencies are highest in the climatological jet exit regions located over the North Pacific and North Atlantic oceanic basins.

Annually, AWB is more frequent than CWB (Fig. 4), a result supported by Barnes and Hartmann (2012). This feature in the climatology has been attributed to several possible mechanisms, including but not limited to the following:

1) A preferred mode of AWB due to asymmetries in the refractive index of Rossby waves that favors wave propagation to the anticyclonically sheared southern flank of the jet (e.g., Nakamura 1993; Orlanski 2003; Riviè re 2009).
2) Subtropical monsoon ridges providing a focusing mechanism for downstream AWB (e.g., Postel and Hitchman 1999; Scott and Cammas 2002; Abatzoglou and Magnusdottir 2006; Hitchman and Huesmann 2007).
3) The jet structure in the eastern ocean basins, which often favors AWB between the exit region of the eddy-driven jet and the entrance region of the subtropical jet (Peters and Waugh 2003).
4) The role of the Hadley cell poleward-directed upper-tropospheric branch of high momentum transfer on the equatorward side of the subtropical jet (Barnes and Hartmann 2012).

In the annual mean, the frequency of extratropical RWB is slightly higher in the North Atlantic basin than in the North Pacific basin for both AWB (30% vs 27.5%) and CWB (17.5% vs 12.5%; Fig. 4). The extension of AWB frequency found by Isotta et al. (2008) and Barnes and Hartmann (2012) over continental Europe and Asia is also found in our climatology, although the dominant type of RWB for these regions is AWB for our study (Fig. 4a). We do not capture the relatively high occurrence of CWB in the deep tropics captured by Barnes and Hartmann (2012). These events occur at low latitudes where the DT becomes unreliable and are not included in our climatology. Regardless, our annual
climatology broadly reflects the results of many previous efforts, though with higher frequencies of detection.

We further decompose the climatology of RWB frequencies into seasonal means (Figs. 5, 6). The maximum in AWB frequency translates westward across the North Pacific and North Atlantic basins and increases in frequency during the transition from winter to summer (Fig. 5). North Atlantic AWB is more frequent than North Pacific AWB for winter (30% vs 22.5%) and autumn months (37.5% vs 35%; Figs. 5a,d), while North Pacific AWB exceeds North Atlantic AWB for spring (27.5% vs 25%) and summer months (40% vs 35%; Figs. 5b,c). Areas over central Asia (30°N, 90°E) experience more frequent AWB in the transition seasons (MAM: 12.5%, SON: 15%; Fig. 5), compared to winter (10%) and summer (5%). Cyclonic wave breaking frequencies are highest in the winter and lowest in the summer (Fig. 6). During spring, summer, and autumn months, the North Atlantic basin generally exceeds the North Pacific in CWB frequency, though a maximum in CWB is found over eastern Asia during the summer (Figs. 6b–d). North Pacific CWB shifts westward from a winter maximum of 22.5% near 50°N, 170°W, to a summer maximum of 10% over eastern Asia (55°N, 120°E). The frequency of CWB in the North Pacific is slightly higher in the spring (15%; 55°N, 180°W; Figs. 6b,d). The seasonal peak in North Atlantic CWB is found near 55°N, 45°W, for winter (22.5%), spring (17.5%), and autumn (17.5%), while summer exhibits an extended region of CWB frequencies of 7.5% from 60°W to 40°E near 60°N (Fig. 6).

The seasonal variability of AWB shown by our approach largely captures the variability found in part by several previous studies. The seasonal maximum in AWB over the summer North Pacific basin has been found in previous climatologies (Postel and Hitchman 1999; Scott and Cammas 2002; Abatzoglou and Magnusdottir 2006). The higher frequency of summer AWB events in both basins has been linked to the Asian and North American monsoons (Postel and Hitchman 2001; Abatzoglou and Magnusdottir 2006). Poleward shifts in the jet from winter to summer are also associated with increased frequencies of AWB and decreased frequencies of CWB (e.g., Akahori and Yoden 1997; Rivière 2009). More frequent winter AWB in the North Atlantic basin, compared to the North Pacific basin, has been shown by Martius et al. (2007) and Strong and Magnusdottir (2008b).
but differs from Abatzoglou and Magnusdottir (2006). The increased frequency of North Atlantic AWB in winter has been associated with a preferential split jet stream in the North Atlantic, compared to a single subtropical jet in the North Pacific (Martius et al. 2007). Higher frequencies of AWB in the autumn relative to spring have also been found by Abatzoglou and Magnusdottir (2006), who found that the autumn critical latitude and jet stream was displaced 10° poleward of its spring counterpart, resulting in a more favorable environment for AWB. In addition, peaks in AWB over eastern Asia in spring and autumn likely occur in
response to changes in the mean location of jet stream entrance and exit regions over central Asia (Abatzoglou and Magnusdottir 2006).

Our study also finds general agreement with previous CWB climatologies. Strong and Magnusdottir (2008b) found three maxima in CWB: North Atlantic, North Pacific, and north of India. Our technique finds all except the latter, which may be due to differences in the ability of the techniques to identify events in regions of high terrain such as the Himalayan Mountains. The locations of the regions of maxima in DJF CWB shown here also match well with Woollings et al. (2008) and Gabriel and Peters (2008). Furthermore, the magnitudes in winter CWB frequency over the North Pacific and
North Atlantic found here are similar to those of Martius et al. (2007). Wernli and Sprenger (2007) and Isotta et al. (2008) found a broad poleward contraction and decrease in the frequency of potential vorticity streamers on the 320–330-K surfaces commonly associated with CWB when transitioning from winter to summer. Their locations of highest frequencies differ from ours, however, which is likely due to differences in the identification techniques applied. Hitchman and Huesmann (2007) found more than a 50% frequency of occurrence of potential vorticity reversals on the 320–340-K surfaces in the high latitudes during summer. However, this region was also dominated by a high frequency of cutoff potential vorticity features that could be contributed by both RWB and other cutoff-type features such as tropopause polar vortices.

Our approach toward identifying RWB events on the DT generally captures the distributions of many previous climatologies with respect to the locations of RWB occurrence as well as with many of the differences between seasons. However, our technique also finds frequencies of occurrence that generally exceed those of any other prior study. We believe that the higher frequencies of occurrence are primarily a function of our identification scheme that identifies a much larger region to be undergoing RWB, which differs from previously implemented methods (Fig. 7). Other differences in frequencies may lie in (i) how a particular method constrains the temporal duration of RWB (e.g., instantaneous vs longer duration), (ii) the surface(s) utilized to search for RWB (e.g., single or multiple isentropic, isobaric, or vorticity surfaces), or (iii) the point in the RWB life cycle that an algorithm identifies the onset or completion of RWB.

4. 1980–2010 trends in AWB and CWB

Trends in the frequency of RWB have been previously explored (Isotta et al. 2008; Rivièrè 2011; Barnes and Hartmann 2012). Most of the identified trends have been attributed to changes in regions of strong baroclinicity, which can be realized as shifts in the climatological jet location (leading to changes in the frequency of RWB; Barnes and Hartmann 2012), or changes in the eddy length scale that can in turn modulate the frequency of RWB (resulting in shifts in the climatological jet location; Rivièrè 2011). While seasonal trends in stratospheric and tropospheric potential vorticity streamers have been diagnosed on isentropic surfaces from 1958 to 2002 in the ERA-40 dataset (Isotta et al. 2008), a strict decomposition of AWB and CWB events has not been performed, nor has such an analysis been performed in the NCEP Reanalysis-2 dataset. Here, we present seasonal trends in Northern Hemisphere AWB and CWB for the 1980–2010 period (Figs. 8, 9), along with trends in the zonal (u) and meridional (v) components of the DT wind (Figs. 10, 11). Trends in each field are calculated using a simple linear regression of seasonal-mean RWB frequencies for each grid point in the hemisphere.
The North Pacific basin is subject to a westward shift in the climatological maxima in AWB for DJF with a 0.4% yr\(^{-1}\) trend near 30\(^\circ\)N, 150\(^\circ\)W, and a \(-0.3\%\) yr\(^{-1}\) trend near 30\(^\circ\)N, 110\(^\circ\)W (Fig. 8a). A significant decreasing trend (\(-0.2\%\) yr\(^{-1}\)) in CWB is also found for the North Pacific basin (Fig. 9a). These trends in North Pacific AWB are found to be associated with a poleward-shifted jet, indicated by the south–north dipolar anomaly in zonal winds across the North Pacific basin, and a split jet in the eastern North Pacific, indicated by a significant westerly

![Figure 8](image_url)
zonal anomaly near 20° and 60°N (Fig. 10a). Changes in the meridional wind are suggestive of those that could be found for AWB, with an equatorward-tilted region of anomalous southerlies (35°N, 180°), northerlies (35°N, 145°W), and southerlies (20°N, 130°W; Fig. 11a).

Positive trends in MAM AWB are found over the Gulf of Alaska, Bering Sea, Gulf of Mexico, and east coast of North America (Fig. 8b). The southern periphery of the climatological CWB region in the North Pacific (40°N, 180°–140°W) is subject to negative trends during this period, while the western flank of the North Atlantic CWB region has a positive trend in CWB (Fig. 9b). These trends can likely be associated with localized poleward shifts in the jet over the Bering.
Sea (nonsignificant) and North America (significant), while the exit region of the North Atlantic jet is equatorward shifted (Fig. 10b). Summer trends in AWB show a westward trend over the western and central North Pacific basin along with a positive trend in CWB near Japan (Figs. 8c, 9c). Further, a positive trend in AWB is noted near 60°–80°N, 60°E (Fig. 8c). A clear attribution is difficult to ascertain from the trends.

Fig. 10. The 1980–2010 linear trend in the Northern Hemisphere DT (2.0-PVU surface) zonal wind (\(\bar{u}_{DT}\)) by season (shaded; kt yr\(^{-1}\); 1 kt \(\approx 0.5144\) m s\(^{-1}\)): (a) winter (DJF), (b) spring (MAM), (c) summer (JJA), and (d) autumn (SON). Seasonal climatological-mean \(\bar{u}_{DT}\) is contoured (black contour) every 20 kt (starting from 20 kt) for positive values and contoured (black dashed contour) every \(-20\) kt (starting from \(-10\) kt) for negative values. Stippling indicates regions of linear trends of \(\bar{u}_{DT}\) with \(p\) values less than 0.05.
in $u_{DT}$ and $v_{DT}$, though a localized poleward shift in the jet over East Asia is noted near 100°E (Fig. 10c). However, an apparent enhancement of a Rossby wave train-like pattern over Europe and Asia (30°–60°N, 0°–120°E; Fig. 11c) may play an important role in seeding the North Pacific jet with enhanced Rossby wave activity.

Trends in autumn (SON) AWB are broadly negative along the southern and southeastern periphery of both the North Pacific (up to $-0.3\% \text{yr}^{-1}$ along the 15°–20°N band)
and North Atlantic (up to −0.3% yr⁻¹ along the 20°N band) climatological-mean AWB regions (Fig. 8d). A positive trend in AWB is also found near 35°N, 80°E (Fig. 8d). CWB trends are positive for both basins as well, with regions of significant positive trends from 40° to 80°N near 160°W in the North Pacific and near 45°N from 40° to 60°W in the North Atlantic (Fig. 9d). Interestingly, significant trends in the North Pacific jet from 120°E to 180° and the subtropical jet over North Africa indicate a poleward-shifted jet (Fig. 10d). Southern portions of the subtropical jet entrance regions in both basins may also be exhibiting a poleward displacement with anomalous slowing in the band from 0° to 20°N in the Western Hemisphere. Such a poleward shift in the jet would be expected to favor increases in AWB and decreases in CWB. However, there is evidence of an enhanced Rossby wave train with poleward propagation over the North Atlantic that decays near the Bering Sea that may act to enhance CWB in the North Pacific while decreasing AWB in the southern and southeastern regions of both basins (Fig. 11d).

Positive trends in North Atlantic and North Pacific AWB in DJF and MAM, negative trends in North Atlantic AWB for JJA, and negative trends in eastern North Pacific AWB for SON (Fig. 8) match well to trends in potential vorticity streamers on the 340–350-K isentropes shown by Isotta et al. (2008, their Figs. 2 and 3). Further agreement is found for CWB, where negative trends in the North Pacific and positive trends in the North Atlantic for MAM, and positive trends in the North Pacific for SON (Figs. 9b,d), match well to their identified trends on the 320–330-K surfaces. Several differences are identified from Isotta et al. (2008), however, including negative trends in North Pacific DJF CWB, increased and westward trends in North Pacific JJA AWB, and negative trends in North Atlantic SON AWB (Figs. 8c,d, 9a). These differences could be due to a variety of physical processes or differences in the reanalysis sets and merit future investigation. Further, these changes in the frequency of RWB can play an important role in modulating momentum transfer into the eddy-driven jet. Such changes can manifest as variability in the structure of the eddy-driven jet in the present and future climate (e.g., Rivière 2011), making them an interesting area of continued research.

5. Interannual variability of AWB and CWB

The relative frequencies of RWB for different ocean basins have been explored in relation to patterns of low-frequency variability, including the NAO (e.g., Benedict et al. 2004; Martius et al. 2007; Rivière and Orlanski 2007; Strong and Magnusdottir 2008b), ENSO (e.g., Waugh and Polvani 2000; Shapiro et al. 2001; Rivière and Orlanski 2007), and PNA (e.g., Orlanski 2005; Martius et al. 2007; Franzke et al. 2011). Here, we present the interannual variability of seasonal-mean Northern Hemisphere RWB, examine correlations between regional RWB and seasonal-mean teleconnections, and explore correlations between AWB and CWB in an individual basin and AWB or CWB between basins.

To first examine the seasonal variability of RWB, longitudinally binned RWB frequencies for 1980–2010 are shown for latitudinal bands that approximately confine the climatological-mean regions of AWB (20°–60°N) and CWB (40°–80°N; Fig. 12). Identified trends in the climatology of AWB identified in Fig. 8 are also evident here, including the westward shift of North Pacific AWB in DJF (Figs. 8a, 12a) and the increased frequency of North Pacific AWB in MAM (Figs. 8b, 12c). These differences could be due to a variety of phenomena, including the transition from the positive to the negative phase of the PDO (Strong and Magnusdottir 2009).

Notable periods of interannual variability are present in all seasons for both AWB and CWB (Fig. 12). A temporal oscillation of anomalies on the order of every 4–6 years is notable in DJF for AWB and CWB and in JJA for AWB (Figs. 12a,b,e). Shifts in the location of wave breaking anomalies within a respective basin are also observed. An example of this can be found for DJF AWB in the North Atlantic region, where a western positive anomaly is found from 30° to 60°W in 1993/94, an eastern positive anomaly from 0° to 30°E in 1991/92, and a split in positive anomalies in 2007/08. These differences could be due to a variety of influences, including periods of low-frequency variability influencing the locations of the jet stream and low-level baroclinicity, that warrant future research.

The role of RWB in establishing teleconnections, or vice versa, has been well established in the literature. Here, we examine the correlations (Table 1) between the NAO, EP, PNA, ENSO-3.4, and AO seasonal-mean indices and (i) AWB in the North Pacific (AWBNP), (ii) CWB in the North Pacific (CWBNP), (iii) AWB in the North Atlantic (AWBNA), and (iv) CWB in the North Atlantic (CWBNA). The domains for each of these four regions are summarized in Table 1. The temporal variability of select teleconnections and RWB frequencies are further shown in Fig. 13. The PNA and ENSO are significantly anticorrelated with AWBNP for all seasons except JJA and SON, respectively (Table 1). Significant correlations are found between CWBNP and the EP in all seasons except MAM, and the PNA in DJF and SON (Table 1). The correlations between the DJF PNA and AWBNP and CWBNP correspond with those found by Franzke et al. (2011).
FIG. 12. The 1979–2011 seasonal-mean (left) AWB and (right) CWB detection frequency by longitude. Seasonal-mean RWB frequencies are calculated as the seasonal-mean frequency (%; shaded) from 20° to 60°N for AWB and 40° to 80°N for CWB at each longitudinal band (2.5°). Seasonal departures of the climatological mean (1980–2010) are contoured (every 5% excluding 0%; positive: red, negative: blue).
The NAO and AWBNA exhibit a significant positive correlation in all seasons except JJA, while CWBNA has a significant negative correlation with the NAO and positive correlations with the PNA and ENSO in DJF. The role of AWBNA in establishing the phase of AWB or CWB between basins that pass our significance test (p value 0.05), though DJF AWBNP and AWBNA are strongly positively correlated (p value 0.055). This strongly correlated value is not surprising, given the established role in eastern North Pacific AWB events in helping to trigger downstream North Atlantic AWB in the winter season, which also helps to strongly modulate the NAO (Benedict et al. 2004; Strong and Magnusdottir 2008b; Rivière and Drouard 2015a,b). Next, we examine select hemispheric variables associated with periods of intrabasin high frequencies of AWB and low AWB in winter. To do so, we examine the 1988/89 winter season, which had high AWB in the North Atlantic and North Pacific basins; the 2009/10 winter season, which had low AWB in both basins; and

The NAO and AWBNA exhibit a significant positive correlation in all seasons except JJA, while CWBNA has a significant negative correlation with the NAO and positive correlations with the PNA and ENSO in DJF. The role of AWBNA in establishing the phase of the NAO has been well established (e.g., Benedict et al. 2004; Martius et al. 2007; Strong and Magnusdottir 2008b) and matches well to our results. Last, the AO significantly correlates with all wave breaking types in the winter season and demonstrates a general positive correlation with AWB and negative correlation with CWB (Table 1). This overlap between significant correlations for the AO, NAO, and EP is not entirely surprising, as all three can be considered a function of the NAM, which has been tied to RWB frequencies previously by Drouard et al. (2015).

Despite the established close link between the jet position and frequencies of CWB and AWB, only the AO and the PNA in DJF and the PNA in SON have significant correlations for both CWB and AWBNP. While the AO and NAO in DJF are the only teleconnections to significantly correlate to both CWB and AWB (Table 1). When correlating the frequency of seasonal RWB within a specific basin, AWBNP and CWB show significant anticorrelations in DJF and SON, while AWBNP and CWB are significantly anticorrelated in DJF (Table 2). These correlations indicate the complicated nonlinear interactions among low-frequency variability, jet stream position, momentum transfer, and RWB, among other features, which can alter the interactions between AWB and CWB in a given region.

In addition to seasonal variability of RWB within a specific basin, the importance of RWB in one basin influencing that within another holds important information for establishing low-frequency patterns that may influence predictability and teleconnection patterns (e.g., Benedict et al. 2004; Drouard et al. 2015; Rivière and Drouard 2015a). Interesting periods of coupled basin variability are noted in Fig. 12a, including periods of high (1988/89, 2004/05) and low (1997/98, 2009/10) DJF AWBNP and AWBNA, among other seasons and RWB types. To investigate further, we first examine the correlation between basins for seasonal AWB and CWB frequencies (Table 2). No seasons have correlations of AWB or CWB between basins that pass our significance test (p value <0.05), though DJF AWBNP and AWBNA are strongly positively correlated (p value = 0.055). This strongly correlated value is not surprising, given the established role in eastern North Pacific AWB events in helping to trigger downstream North Atlantic AWB in the winter season, which also helps to strongly modulate the NAO (Benedict et al. 2004; Strong and Magnusdottir 2008b; Rivière and Drouard 2015a,b). Next, we examine select hemispheric variables associated with periods of intrabasin high frequencies of AWB and low AWB in winter. To do so, we examine the 1988/89 winter season, which had high AWB in the North Atlantic and North Pacific basins; the 2009/10 winter season, which had low AWB in both basins; and

### Table 1. Correlation coefficients of the normalized seasonal frequency of identified AWB and CWB with the normalized seasonal-mean NAO, EP, PNA, ENSO 3.4, and AO teleconnection indices. For CWB, the North Pacific region (CWBNP) is defined as 40°–80°N, 160°E–140°W, and the North Atlantic region (CWBNA) is defined as 40°–80°N, 70°–30°W, for all seasons. For AWB, the North Pacific region (AWBNP) is latitudinally confined between 20° and 60°N and longitudinally between 150° and 90°W for DJF; between 180° and 120°W for MAM, 140°E and 160°W for JJA; and between 180° and 120°W for SON. The North Atlantic AWB region (AWBNA) is latitudinally confined between 20° and 60°N and longitudinally between 40°W and 20°E for DJF; between 50°W and 10°E for MAM; between 90° and 30°W for JJA; and between 60°W and 0° for SON. Boldface values signify correlations with a p value less than 0.05.

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<th>PNA</th>
<th>ENSO</th>
<th>AO</th>
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<td></td>
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<td></td>
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FIG. 13. The 1980–2010 time series of seasonal-mean basin RWB frequency (solid) and seasonal-mean normalized teleconnection index (dashed) for (a) DJF PNA and AWB$_{NP}$, (b) DJF PNA and CWB$_{NP}$, (c) SON PNA and AWB$_{NP}$, (d) SON PNA and CWB$_{NP}$, (e) DJF NAO and AWB$_{NA}$, (f) DJF NAO and CWB$_{NA}$, (g) DJF AO and AWB$_{NA}$, (h) DJF AO and CWB$_{NA}$, (i) DJF AO and AWB$_{NP}$, and (j) DJF AO and CWB$_{NP}$. All time series are correlated or anticorrelated with $p$ values less than 0.05 and are summarized in Table 1.
TABLE 2. Correlation coefficients between normalized seasonal-mean AWB and CWB frequency within the North Pacific and North Atlantic basin regions (columns 2 and 3) and between the North Atlantic and North Pacific basin regions for seasonal-mean AWB and CWB frequency, respectively (columns 4 and 5). The AWBNP, AWBNA, CWBNP, and CWBNA regions are as defined in Table 1. Boldface values signify correlations with a p value less than 0.05.

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6. Conclusions

We have introduced a new approach for cataloging RWB events in all seasons on the DT. Instantaneous AWB and CWB were identified using a contour identification technique similar to that of Barnes and Hartmann (2012) for 1979–2011 in the 6-hourly NCEP Reanalysis-2 global reanalysis dataset for the Northern Hemisphere. Overturning contours of potential temperature are identified on the 2.0-PVU surface that is used to approximate the DT (Hoskins et al. 1985).

Rossby wave breaking events were identified and cataloged in a manner that captures both the poleward and equatorward high and low potential temperature streamers as part of a singular RWB event. This approach toward identifying regions of RWB represents a departure from previous studies that have typically identified one streamer or the other or the potential temperature (or potential vorticity) gradient reversal between the two. Case studies of two RWB events (one AWB event and one CWB event) demonstrated (i) the importance of capturing both regions of an RWB event when considering the dynamic and thermodynamic perturbations through the depth of the troposphere and (ii) the effectiveness of our RWB region method in capturing the majority of the localized atmospheric response.

The climatology of CWB and AWB presented in this paper encompasses many of the previously shown climatologies that were based on a variety of identification schemes. Our approach is not meant to replace previous techniques, but to instead provide an alternative approach toward the automated identification of these phenomena that avoids some of the potential limitations of other techniques. Differences in the climatologies can generally be explained by differences in the identification methods applied. For example, our technique identifies the RWB region, which encompasses the full wavelength of the RWB as opposed to a single streamer or the region of potential temperature (or potential vorticity) gradient reversal, resulting in a higher frequency of detection. In doing so, we identify more of the region that we consider to be of importance when

...the difference between the two seasons (Fig. 14). In the 1988/89 season, a depressed tropopause near 20°N, 160°–120°W, results in an enhanced jet entrance region in the low latitudes and a poleward-shifted North Pacific jet exit region (Figs. 14a,d). A similar but less pronounced anomaly is also present in the eastern North Atlantic basin. The meridional wind on the DT is suggestive of what would be expected for an AWB pattern (similar to Fig. 11a) in the eastern North Pacific (Fig. 14g). Thickness anomalies over western Europe indicate a weakening of the lower-tropospheric baroclinic zone, which can occur in association with AWB (Fig. 14j). For 2009/10, the North Atlantic jet is notably elongated with equatorward shifted exit regions in both the North Pacific and North Atlantic in association with an enhanced pressure gradient on the DT (Figs. 14b,e). Such an equatorward shift would be expected to be associated with reduced AWB frequencies (e.g., Rivière 2009). Anomalously large 1000–500-hPa thicknesses are noted over the high latitudes in the Western Hemisphere and an equatorward shifted zonal band of anomalous baroclinicity near 20°–25°N (Fig. 14k).

The differences in the location of AWB frequencies between these two seasons are likely due to, or are a symptom of, the complex nonlinear interactions between RWB and the jet stream (Fig. 14f), which are largely driven by differences in the orientation of the lower-tropospheric baroclinic zones (Fig. 14i). Furthermore, the 1988/89 season was characterized by positive phases of the AO and NAO and a negative phase of the PNA, while the 2009/10 season featured negative phases of the AO and NAO and a positive phase of the PNA (Figs. 12a,g). This result closely parallels those of Drouard et al. (2015), who demonstrated that positive phases of the NAM, which are associated with a poleward displaced jet, are associated with increased AWB and decreased CWB in both the North Atlantic and North Pacific basins. The importance of multiple patterns of low-frequency variability influencing the frequency of RWB, or the opposite, remains an area of continued research (e.g., Shapiro et al. 2001; Martius et al. 2007; Strong and Magnusdottir 2009; Michel and Rivière 2011; Drouard et al. 2015; Rivière and Drouard 2015b; Martius and Rivière 2016) and holds key information for extended predictability (e.g., Benedict et al. 2004).
FIG. 14. DJF mean tropospheric fields for (left) 1989/90, (middle) 2009/10, and (right) 1989/90 – 2009/10 difference. Plotted in black contours are seasonal-mean DT (2.0-PVU surface) pressure (hPa; 20-hPa contour interval), $u_{DT}$ (kt; 20-kt contour interval from 30 kt), and $v_{DT}$ (kt; 5-kt contour interval; dashed negative) in first three rows, and 1000–500-hPa thickness (m; 60-m contour interval) in bottom row. Shading in the left and center columns are the variable anomalies from the 1980–2010 DJF climatology. (right) The 1989–90 DJF mean fields are contoured, and the difference between the 1989/90 and 2009/10 seasonal-mean fields are shaded for each variable.
considering the tropospheric features commonly associated with CWB and AWB.

We have presented a technique that identifies the majority of the areas of localized dynamical importance during an RWB event that can be applied to a wide spectrum of RWB-related investigations. A companion paper (Bowley et al. 2019) examines the specific role of AWB in modulating Northern Hemisphere available potential energy. It is important to note, however, that other techniques may still be better applied when isolating specific phenomena. For example, when examining blocking phenomena, a technique that specifies the low potential vorticity streamer may be most appropriate, while studies examining extreme cyclones may instead focus entirely on the high potential vorticity streamer. Therefore, it is crucial that future studies consider the technique that is most appropriate based upon their atmospheric phenomena of interest. Further work could be performed using this technique, including examining changes in the latitudinal and longitudinal extent of RWB events and the magnitude of the RWB event (i.e., how many isentropes break). These properties may influence high-impact phenomena such as atmospheric rivers and blocking, the latter of which have been shown to be strongly forced by RWB (Masato et al. 2012, 2013). Finally, testing the same technique across a series of reanalysis datasets could provide interesting insight into the relative frequencies of RWB in each dataset, as well as examine the role of resolution in the RWB identification scheme.

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REFERENCES


