A Climatology of Rossby Wave Breaking on the Southern Hemisphere Tropopause

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ABSTRACT

A 30-yr climatology of Rossby wave breaking (RWB) on the Southern Hemisphere (SH) tropopause is formed using 30 yr of reanalyses. Composite analysis of potential vorticity and meridional fluxes of wave activity show that RWB in the SH can be divided into two broad categories: anticyclonic and cyclonic events. While there is only weak asymmetry in the meridional direction and most events cannot be classified as equatorward or poleward in terms of the potential vorticity structure, the position and structure of the fluxes associated with equatorward breaking differs from those of poleward breaking. Anticyclonic breaking is more common than cyclonic breaking, except on the lower isentrope examined (320 K). There are marked differences in the seasonal variations of RWB on the two surfaces, with a winter minimum for RWB around 350 K but a summer minimum for RWB around 330 K. These seasonal variations are due to changes in the location of the tropospheric jets and dynamical tropopause. During winter the subtropical jet and tropopause at 350 K are collocated in the Australian–South Pacific Ocean region, resulting in a seasonal minimum in the 350-K RWB. During summer the polar front jet and 330-K tropopause are collocated over the Southern Atlantic and Indian Oceans, inhibiting RWB in this region.

1. Introduction

The propagation and breaking of Rossby waves near the tropopause plays an important role in the atmospheric general circulation and in determining the distribution of trace gases. In particular, Rossby wave breaking (RWB), which involves the irreversible deformation of potential vorticity (PV) contours (e.g., McIntyre and Palmer 1983, 1984), has been shown to be important for understanding a wide range of atmospheric processes, including transport between the stratosphere and troposphere (e.g., Scott and Cammas 2002; Sprenger et al. 2007), atmospheric blocking (Pelly and Hoskins 2003; Berrisford et al. 2007), cutoff low pressure systems (Baray et al. 2003; Ndarana and Waugh 2010), and annular modes (Benedict et al. 2004; Rivière and Orlanski 2007; Strong and Magnusdottir 2008; Woollings et al. 2008).

Given this importance there is an extensive literature examining the occurrence, evolution, and cause of upper-tropospheric RWB, using both observations and models. These studies have identified four paradigms for the morphology of RWB, depending on the barotropic shear, the sense of rotation of PV features (“anticyclonic” or “cyclonic”), and the predominant direction of the advection of air (“equatorward” if event involves mainly advection of high-PV air toward the equator or “poleward” if advection of low-PV air toward the pole) (e.g., Esler and Haynes 1999; Gabriel and Peters 2008, hereafter GP08). Thornicroft et al. (1993) first contrasted the behavior of equatorward RWB in anticyclonic and cyclonic shear. These two types of RWB are now generally referred to as LC1 and LC2 events, respectively. Peters and Waugh (1996) then provided a similar characterization of poleward RWB in cyclonic and anticyclonic shear (P1 and P2 events, respectively). The classification into these four types is useful for understanding the possible impact of RWB on the general circulation, weather
patterns, and exchange between the stratosphere and troposphere. For example, LC1 and P2 events are both anticyclonic, but they are connected with very different weather patterns and stratosphere–troposphere exchange: LC1 events are associated with thin filaments of high PV that produce cutoff cyclones and transport from the stratosphere into the troposphere, whereas P2 events are associated with large regions of anomalously low PV, blocking anticyclones, and transport into the stratosphere.

Several subsequent observational and modeling studies have examined the cause and occurrence of these types of RWB (e.g., Hartmann 1995; Hartmann and Zuercher 1998; Hartmann 2000; Peters and Waugh 2003; Moon and Feldstein 2009; Wittman et al. 2007; Kunz et al. 2009), with most studies focusing on either equatorward or poleward events. One exception is the study of Esler and Haynes (1999), who showed that all four paradigms occurred in an idealized model of the tropospheric circulation. More recently, GP08 examined the climatological occurrence of the four types of wave breaking in the Northern Hemisphere, using 45 yr of meteorological reanalyses.

A similar analysis has not been performed for the Southern Hemisphere (SH), and one objective of this study is to examine the RWB morphology and occurrence of the four paradigms in the SH. There is currently no consensus on the extent that breaking SH can be separated into equatorward and poleward events. Peters and Waugh (2003) showed that some basic-state wind configurations can lead to equatorward breaking and others to poleward advection of tropospheric air, implying that poleward/equatorward asymmetry exists for RWB. However, Berrisford et al. (2007) considered RWB that is associated with blocking and suggested that RWB in the SH is either cyclonic or anticyclonic but symmetric in the meridional direction. We address this here by using an approach similar to that of GP08 to examine the occurrence and morphology of the four types of breaking in the SH, using meteorological reanalyses for 1979–2008.

A second objective is to determine the seasonal and spatial variations of RWB in the SH and the relationship to the seasonal march of the tropospheric jets. This is done by forming climatologies of RWB, zonal winds, and the dynamical tropopause. Several previous studies have formed climatologies of RWB in SH (e.g., Postel and Hitchman 1999; Hitchman and Huesmann 2007; Berrisford et al. 2007), but these have not focused on the different types of wave breaking.

The structure of the paper is as follows: in the next section data and methods are outlined, the climatological structures of the different types of RWB are examined in section 3, seasonal and spatial variations are considered in section 4, and in section 5 we give the summary and conclusions.

2. Data and methods

a. Data

We use the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis fields (Kalnay et al. 1996) to examine RWB events and associated processes in the SH from 1979 to 2008. The horizontal wind, temperature, and geopotential heights data are on a grid with a horizontal resolution of 2.5° × 2.5° on all tropospheric standard levels.

The PV expression used is given by

\[ P = \frac{g}{\cos \theta} \left[ f + \frac{1}{a \cos \phi} \frac{\partial v}{\partial \lambda} - \frac{1}{a \cos \phi} \frac{\partial (u \cos \phi)}{\partial \phi} \right] \left( \frac{\partial \theta}{\partial \rho} \right), \]

where \( u \) and \( v \) are the zonal and meridional velocity fields, respectively, \( \theta \) is the potential temperature, \( f \) is the Coriolis parameter, and \( a \) is the radius of the earth (see Postel and Hitchman 1999). The stability \( -\partial \rho / \partial \theta \) is calculated in terms of temperature as in Trenberth (1991). PV is calculated on isobaric surfaces from daily velocity and temperature fields. The spatial derivatives of the velocity fields are calculated using second-order centered differences. The resulting isobaric PV fields are then linearly interpolated to isentropic surfaces, as in Edouard et al. (1997).

The quality of NCEP–NCAR reanalysis in the SH is questionable prior to 1979 since it was poorly observed during this era because it is mostly covered by oceans (Tennant 2004). Improvements to this state of affairs were effected by the introduction of Television and Infrared Observation Satellite (TIROS) Operational Vertical Sounder (TOVS) in 1979 (Kistler et al. 2001). Therefore, we restrict our analysis from 1979 onward.

b. RWB identification

We focus here on RWB events that occur along the tropopause. We identify RWB events on PV = \(-1.5, -2.0, \) and \(-2.5 \) PVU (1 PVU = \( 10^6 \) K m² s⁻¹ kg⁻¹) contours on the 320-, 330-, 340-, and 350-K isentropic surfaces. These contours are located along the tight meridional PV gradient that approximates the dynamical tropopause. The intersection of the dynamical tropopause (PV = \(-1.5\) to \(-2.5\)-PVU contours) and isentropes occurs at higher latitudes for lower isentropes and also varies with season (see Fig. 1). For the isentropes considered here, the intersection (and hence RWB) varies from the subtropics to middle latitudes.
To identify the RWB events, we employ an objective algorithm that uses the geometry of the overturning contours that represents the RWB process (e.g., Esler and Haynes 1999; Postel and Hitchman 1999). This is a multi-step process. First, subsynoptic-scale features are filtered out from the PV by spectral truncation, leaving only the zonal mean and the first 10 wavenumbers. Then on each day and for each of the three PV contours, potential RWB events are identified when a meridional line intersects a PV contour at two or more latitudes (cf. Fig. 5a in Esler and Haynes 1999). When these intersections are found, only those grid points for which \( \psi_y \), 0 (Baldwin and Holton 1988; Postel and Hitchman 1999; Hitchman and Huesmann 2007) are considered to belong to a RWB event. This ensures the exclusion of intrusions that occur with no overturning. If grid points that are identified in this manner have equal or consecutive longitudes, then they belong to the same RWB event and so only the time, longitude, and latitude of the westernmost grid point that satisfies this gradient reversal condition are recorded.

This method inherently captures closed contours that represent secluded stratospheric and tropospheric air blobs (closed high and low PV anomalies). To exclude such features from the database, we apply it only to contours that have more than 144 grid points; this ensures that each contour that we consider is as long as a latitude circle or longer. The blobs are likely to be residues of old RWB events [see schematics in Thorncroft et al. (1993) and Peters and Waugh (1996)], so we remove them here to avoid duplicate counts of events. In some cases, the same RWB event occurs on two or all three contours simultaneously. In such cases only one of the points is recorded. Finally, to ensure that each RWB event is accounted for only once, the algorithm keeps only the count that is associated with the onset of the event. The resulting database contains the dates, longitudes, latitudes, and values of the PV contour for the onset of RWB events. For the isentropes considered here the intersection, and hence RWB, varies from latitudes around 25°S to 55°S. (This range is larger than shown in Fig. 1 because we consider PV = −2 PVU and −2.5 PVU as well as −2 PVU and because there can be large meridional variations in the location of the PV contours for any given day and longitude.)

c. Separating RWB into different types

The above algorithm does not differentiate between the types of RWB events that were discussed in the introduction. To divide our database into these categories, we use a two-step scheme that closely follows that of GP08. In the first step, the algorithm divides the events into cyclonic and anticyclonic types using the meridional component \( F^y \) of the flux vector of wave activity for zonally asymmetric basic-state flows. The wave activity density associated with this meridional flux is a linear combination of wave energy and potential enstrophy and obeys a conservation law (Takaya and Nakamura 2001). The form of the meridional flux component implemented here is given by

\[
F^y = \frac{\cos \phi}{2|u_o|} \left[ u_o \left( -u'v' + \frac{\psi'}{a \cos \phi} \frac{\partial u'}{\partial \lambda} \right) + v_o \left( u'^2 + \frac{\psi'}{a \cos \phi} \frac{\partial u'}{\partial \phi} \right) \right],
\]

(2)

where \( u, v, \) and \( a \) are defined as in (1) and \( \psi \) is the geostrophic streamfunction. Here \( u' \) corresponds to

![Fig. 1. Zonally averaged winds (thin contours) for (a) DJF and (b) JJA with isentropic (thick black) contours and the PV = −2 PVU (thick gray) contour approximating the dynamical tropopause. The zonal winds are drawn at 5 m s\(^{-1}\) contour intervals and the isentropes at 10-K intervals.](image-url)
perturbation velocity and is defined as $u' = u - u_0$, where $u_0$ is the basic state zonal velocity; $v'$ and $\psi'$ are defined in a similar manner. Esler and Haynes (1999) were the first to demonstrate the use of such wave activity diagnostics to study the characteristics of RWB. They defined a wave activity flux index, which is a function of longitude and time, in terms of $F^y$ and then used it to show that poleward and equatorward fluxes are associated with cyclonic and anticyclonic RWB, respectively. These two types of breaking will be respectively referred to as CWB and AWB hereafter. The relationship between the type of RWB and the direction of wave activity is facilitated by the fact that the sign of the meridional wavenumber in linear wave theory determines the meridional direction in which the flux propagates (see section 4.5.5 in Andrews et al. 1987). The equatorward/westward–poleward-eastward (poleward/ westward–equatorward/eastward) tilt of the AWB (CWB) is associated with a positive (negative) meridional wavenumber and hence the equatorward (poleward) flux of wave activity. The difference between Esler and Haynes (1999) and GP08 is that the latter employed the actual wave activity flux index, which is a function of longitude and time, in terms of $F^y$ and then used it to show that poleward and equatorward fluxes are associated with positive (negative) fluxes in the boundary region. The relative magnitude of the mean of the positive and negative fluxes is then used to classify the events as anticyclonic or cyclonic (Esler and Haynes 1999; GP08). If the mean of the positive fluxes is larger (i.e., $F^y_A > F^y_C$), then the RWB event in question is classified as an AWB event, as indicated by the subscript $A$. If the reverse holds, the event is classified as a CWB event. This scheme was applied to all RWB events on all isentropic surfaces. Note that a visual inspection of individual events on the 340- and 350-K isentropic surfaces made it necessary to expand the boundary region as some of the positive fluxes associated with AWB were displaced westward and equatorward of the fold.

The second step of the algorithm uses both the fluxes and the ambient flow conditions to define equatorward–poleward asymmetry of the RWB events (i.e., AWB are divided into LC1 and P2, and CWB events into LC2 and P1). GP08 showed that LC2 and P2 events, which develop downstream and are associated with diverging isolines, occur within diffluent flow, whereas LC1 and P1 are associated with confluent flow (see Fig. 1 of GP08). They further showed that the confluence and diffluence of the flow can be used to separate events into equatorward and poleward events. To identify diffuseness and confluence in the flow field, we use

$$\Phi_{\lambda,\phi} = \frac{1}{a^2 \cos \phi} \frac{\partial^2 \Phi}{\partial \lambda \partial \phi},$$

where $\Phi$ is the geopotential height. In the SH, if $\Phi_{\lambda,\phi} < 0$ ($\Phi_{\lambda,\phi} > 0$) then the flow is diffluent (confluent). As in the first step, we calculate the fields of $\Phi_{\lambda,\phi}$ using centered finite differences. GP08 used 300-hPa geopotential heights for this purpose; in this study we employ the 250-hPa level to accommodate RWB events on both the 350- and 330-K isentropic surfaces.

For the classification of equatorward–poleward asymmetry we consider the AWB and CWB events separately. For AWB events the majority of the fluxes in the boundary region are positive. We then calculate the means of the positive fluxes occurring in confluent and diffluent flow using

$$F^y_{L1} = \frac{1}{N_e} \sum_{e} F^y_e (\Phi_{\lambda,\phi} > 0) \quad \text{and}$$

$$F^y_{P2} = \frac{1}{N_d} \sum_{d} F^y_d (\Phi_{\lambda,\phi} < 0),$$

respectively, where $N_e$ ($N_d$) is the number of grid points at which the fluxes are positive (negative) in the boundary region. The relative magnitude of the mean of the positive and negative fluxes is then used to classify the events as anticyclonic or cyclonic (Esler and Haynes 1999; GP08).
respectively, where $N_c$ ($N_d$) is the number of grid points at which positive fluxes in the boundary region occur in confluent (diffluent) flow. Events are classified as LC1 if $F_{y,L_1}^p$ exceeds $F_{y,P_2}$, and as P2 otherwise. For CWB events, a similar procedure is followed using

$$F_{y,P_1}^c = \frac{1}{N_c} \sum F_{y}^c (\Phi_{\lambda \phi} > 0) \quad \text{and}$$

$$F_{y,L_2}^c = \frac{1}{N_d} \sum F_{y}^c (\Phi_{\lambda \phi} < 0),$$

with events classified as LC2 if $F_{y,L_2}^c$ exceeds $F_{y,P_1}$, and as P1 otherwise.

In summary, a RWB event is classified as

(i) LC1 if $F_{y,A}^c > F_{y,C}^c$ and $F_{y,L_1}^c > F_{y,P_2}$,
(ii) P2 if $F_{y,A}^c > F_{y,C}^c$ and $F_{y,L_1}^c < F_{y,P_2}$,
(iii) LC2 if $F_{y,A}^c < F_{y,C}^c$ and $F_{y,L_2}^c > F_{y,P_1}$, and
(iv) P1 if $F_{y,A}^c < F_{y,C}^c$ and $F_{y,L_2}^c < F_{y,P_1}$.

The separation scheme can then be written in terms of ratios. When implementing this scheme, we remove events with small asymmetry in fluxes given in (5) and (6). Following GP08, for an LC1 event we require the ratio $R_{L_1} = F_{y,L_1}^c/F_{y,P_2}$ to exceed 1.1 and for a P2 event we require $R_{P_2} = F_{y,P_2}^c/F_{y,L_1}^c > 1.1$. The LC2 and P1 events have similar requirements.

d. Composite analysis

To analyze the mean structure of the different types of RWB and their associated meridional fluxes of wave activity, we make use of composite analysis. In such an analysis the fields are shifted in longitude and latitude before averaging so that the centroid points of the RWB events coincide. We use this point as the superposition to avoid the effects of destructive interference and undue smoothing of the larger events on smaller ones. Maps of the composite mean fields of PV and meridional fluxes are then shown in longitude and latitude relative to the location of the centroid points. The composite process is also performed for days before or after the onset of the RWB.

3. Morphology of RWB

Before considering the spatial and temporal occurrence of RWB events in the SH, we first examine the composite mean spatial structure and evolution of the different types of RWB. Figure 2 shows the composite-mean PV = $-2$ PVU contour and $F^y$ fields for AWB and CWB on the 350- and 330-K isentropic surfaces. Similar structures occur on the other isentropic surface. There is a clear difference between the morphology of PV for these two types of RWB. AWB has a northwest–southeast orientation on both surfaces (Figs. 2a,c) while CWB has a southwest–northeast tilt (Figs. 2b,d). Similar structures were observed on the other isentropic surfaces (not shown).

As expected, AWB events are associated with equatorward meridional fluxes of wave activity and poleward

![FIG. 2. (a),(b) Composite mean fields of PV = -2 PVU (thick) contour and meridional component of the flux of wave activity (thin). The PV contours represent AWB on the (a) 350- and (b) 330-K surfaces. (c),(d) As in (a) and (b), respectively, but for CWB. The fluxes are drawn in 10 m$^2$ s$^{-2}$ contour intervals. Positive and negative fluxes are represented by solid and dashed contours, respectively.](attachment://image.png)
Fluxes are seen in the CWB composites (Esler and Haynes 1999). However, the fluxes differ not only in direction but also in strength and structure: Fluxes associated with AWB are stronger than their cyclonic counterparts. The strength of fluxes associated with AWB is suggestive of the propagation of wave activity much farther away from the jet into the subtropics. This wave activity is known to be absorbed in the critical layer region equatorward of the subtropical regions (Held and Hoskins 1985; Randel and Held 1991; Thorncroft et al. 1993). Also, all the fluxes associated with AWB are positive across the domain (Figs. 2a,c). This suggests a dearth of undulations within the wave packet, a physical property that would be expected in the CWB case (Thorncroft et al. 1993). The negative fluxes associated with CWB are flanked by positive ones, suggesting an undulatory behavior (Thorncroft et al. 1993) inside the wave packet, the center of which is where breaking occurs (Lee and Feldstein 1996).

We now consider the difference between equatorward and poleward RWB. Figure 3 shows the evolution of LC1 (left panels) and P2 (right panels). The PV structures and evolution of these two types of RWB are very similar but the differences in their respective fluxes are clear. A visual inspection of individual events indicates that the lack of differences between the PV composites for LC1 and P2 exists because of a combination of the averaging over a large number of events with different PV structures and the fact that many of these structures are symmetric [i.e., there is not a significant difference in characteristics of equatorward (high PV) and poleward (low PV) lobes in most events].

The ratios $R_{LC1}$ and $R_{P2}$ are a measure of equatorward–poleward asymmetry in events and can be used to categorize them; that is, the larger the values of $R_{LC1}$ ($R_{P2}$), the stronger the equatorward (poleward) signature of an AWB event. A similar statement holds for CWB. Figure 4 shows the distribution of the ratios for each type of
RWB. About 15% of events are symmetric with $R_{LC1}$ and $R_{P2}$ ≤ 1.1 (between thick vertical lines in Fig. 4). A visual inspection of individual cases and composite structures of PV and fluxes (not shown) of these events shows that they are quite similar. The latter are equally in confluent and diffuent flow. This symmetry criterion was used by GP08. The similarity in PV extends to cases with ratios greater than 1.1 but the fluxes appear to be slightly distinguishable (not shown), particularly in terms of where they occur relative to the folds. However, these differences are still small, suggesting that poleward–equatorward asymmetry is weak even for values of the ratios between 1.1 and 2.0.

There are some events with large values of the ratios as shown in Fig. 4. Composites of such events are shown in Fig. 5 for the 330-K isentropic surfaces. The differences between the PV structures are slightly clearer for cases with $R_{LC1}$ and $R_{P2}$ greater than 2. While LC1 and P2 are indistinguishable, the differences between the fluxes are much clearer. In the case of LC1 events (Figs. 5a–c), the fluxes occur on the equatorward side of the central latitude, where confluent flow would normally be located. These fluxes have a zonally elongated structure and are stronger than their P2 counterparts from day 0 to day +2. The maximum fluxes occur on the poleward lobe of events where one might expect diffuent flow (see Fig. 1 of GP08).

For the majority of RWB events the PV morphology of LC1 (LC2) is indistinguishable from that of P2 (P1). Therefore, on the one hand RWB events are mostly either cyclonic or anticyclonic with little poleward–equatorward asymmetry in the SH, as suggested by Berrisford et al. (2007). On the other hand, the fluxes associated with LC1 and P2 exhibit pronounced contrasts, with the flux maxima occurring on the equatorward and poleward lobes of the two types of breaking, respectively (cf. Figs. 5a,d).

4. Variations in RWB occurrence

We now examine the temporal and spatial variations in the occurrence of RWB in the SH. Because there are few extreme cases for which there is a clear difference in the PV structure between types of equatorward and poleward breaking, we focus just on the occurrence of AWB or CWB.

a. Seasonal and latitudinal variations

We first discuss the frequency of occurrence and seasonal and latitudinal variations of RWB on the 350- to 320-K surfaces. The method described in section 2 identified a total of 35 973 events on all four isentropic surfaces. There are significant variations of RWB occurrence with isentrope. CWB increases almost fivefold from 350 to 320 K, signaling the gradual importance of cyclonic shear progressively toward the poles. Martius et al. (2007) found a similar type of behavior in the NH. The variations in AWB frequencies appear to be less significant.

Figure 6 shows the frequency of RWB as a function of latitude and month. The seasonal structure of RWB occurrence on the dynamical tropopause varies significantly with isentrope. On the 350-K isentropic surface, both AWB and CWB exhibit a maximum count during December–January (DJF) and a minimum during June–August (JJA), as shown by Postel and Hitchman (1999). A similar behavior is observed on the 340-K surface, with more breaking occurring in late autumn and early
spring. On the 330-K surface, there are fewer events during the summer than on the equatorward surfaces and more during the colder seasons; therefore, the seasonal distribution on the 340-K surface is a transition between the other two surfaces. The situation is slightly different on the 320-K isentropic surface. Here, AWB occurs most frequently during the colder months of the year, while CWB occurs in summer the most.

This figure also shows that the latitudinal position of the intersection of the PV = \(-2\) PVU contour with the each isentropic surface varies with season, as suggested also by Fig. 1. On all surfaces the peak latitude of RWB and the location of the dynamical tropopause are farther south (poleward) in summer than during winter. The most pronounced seasonal shift occurs on the 330- and 320-K surfaces, associated with the steeper dynamical tropopause (Fig. 1; see also Liniger and Davies 2004). The reason for this is the development and intensification of the subtropical jet (STJ), as will be discussed in the next section.

Finally, we note the position of AWB relative to that of CWB events. Using the seasonal position of the dynamical tropopause (which is close to the jet) as a reference, AWB events tend to occur equatorward of their cyclonic counterparts. This is consistent with previous studies (e.g., Thorncroft et al. 1993; Lee and Feldstein 1996; Magnusdottir and Haynes 1996; Peters and Waugh 1996) because AWB is expected to occur on the equatorward side of the jet and CWB on its poleward side.

b. Longitudinal variations

There are also large zonal variations in the occurrence of RWB. This is shown in Figs. 7 and 8, which show maps of the frequency of AWB on the 350- and 330-K surfaces, respectively. Also shown are the zonal winds, represented by the contours of medium thickness, and the 350-K tropopause, shown by the PV = \(-2\) PVU contour (thick, almost zonal contours).

<table>
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<th>TABLE 1. Number of RWB per type.</th>
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On the 350-K surface the variations with longitude differ between seasons. During DJF the zonal variations are small and RWB is common at nearly all longitudes (Fig. 7a). However, during JJA there is a clear peak in RWB over the region covering the western Pacific Ocean, the South Atlantic Ocean, South Africa, and the western Indian Ocean (120°W–90°E), and RWB is rare over the Australia–eastern Pacific Ocean sector (90°E–120°W). As discussed above, there is a strong seasonal variation in RWB at 350 K, with a maximum in DJF and a minimum
in JJA. Figure 7 shows that this seasonal variation is due primarily to variations in the eastern Indian Ocean–western Pacific Ocean sector and there are only weak seasonal variations at other longitudes.

The strong seasonal variations in the eastern Indian Ocean–western Pacific Ocean sector can be understood by the seasonal variations in the position of PV contours approximating the dynamical tropopause and the strength of tropospheric jets. During DJF the subtropical tropopause is away from the jet core and is located in anticyclonic shear at all longitudes, but during the other seasons the tropopause and the jet core are collocated over the eastern Indian Ocean–western Pacific Ocean sector. The pronounced reduction in the number of 350-K RWB events in this sector from DJF to JJA is due to displacement of the PV contours into the regions of the subtropical jet, which develops and attains its maximum in JJA (Hurrell et al. 1998).

Previous studies (e.g., Peters and Waugh 1996; Swanson et al. 1997) have suggested that the waves propagate along strong jets and break in regions of weak zonal flow. Observations of daily maps support this notion. However, these regions of weak zonal flow are few along the STJ, causing the waves to break less often.

The RWB events at 350 K are overwhelmingly anticyclonic during all seasons at all longitudes. The probable reason for this is that most Rossby waves along the chosen PV contours on this surface propagate on the equatorward side of the polar front jet (PFJ), which is located in the midlatitudes in DJF and slightly south during the rest of the year (Hurrell et al. 1998). Therefore, RWB occurrence is likely influenced by anticyclonic shear (Thornicroft et al. 1993; Peters and Waugh 1996, 2003). This confirms that the subtropical events identified by Postel and Hitchman (1999) and Berrisford et al. (2007) during DJF in the middle to western Pacific, South Atlantic, and Indian Oceans are in fact anticyclonic (see Fig. 7). The occurrence of RWB on this surface might provide an additional link between midlatitude baroclinicity and the weather over subtropical countries such as South Africa because the breaking waves are associated with wave activity that originates in the low-level midlatitudes and absorbed just north of 30°S, as noted above. For example, the tropical temperature troughs that are the most important summer rainfall producing systems over this country (van Heerden and Taljaard 1998) are greatly influenced by low-frequency baroclinic eddies (Todd and Washington 1999). Note also

**Fig. 7.** Frequency of AWB (shading separated by dotted contours) on the 350-K surface for (a) DJF, (b) MAM, (c) JJA, and (d) SON. Isotachs (medium thickness) at 250 hPa and the mean position of the dynamical tropopause (PV = -2 PVU) on the 350-K surface (thick contour) are superimposed. The contour interval of the isotachs and RWB frequency are 10 m s⁻¹ and 0.05 events per year.
that CWB in our analysis is rare on this surface because more events of this type would be expected to occur in regions of large $|\text{PV}|$ values, which are in the lower stratosphere as suggested in Fig. 1. The zonal distribution of AWB on the 320-K surface (not shown) is similar to that at 350 K with the exception of occurring slightly poleward.

The seasonal and zonal variations at 330 K differ from those at 350 K, but the same mechanism can still be linked to variations in the jet locations. The seasonal variability of RWB on the 330-K surface is regulated by changes in two sectors, the Southern Ocean (10°W–90°E) and the Australian–western Pacific Ocean region (120°E–120°W). Again, there is a reduction in RWB in these sectors when tropospheric jets are collocated with the dynamical tropopause ($\text{PV} = -2 \text{PVU}$). During DJF the 330-K dynamical tropopause coincides with the PFJ core in the Southern Ocean. At the end of the season, the dynamical tropopause migrates north and remains on the anticyclonic side of this jet right through the rest of the year in most of the SH but is collocated with the subtropical jet in the Australian–western Pacific Ocean region during the wintertime. Note that the more pronounced migration of the 330-K dynamical tropopause discussed above is caused by PV gradients associated with the STJ in this sector. As this jet develops and attains its maximum strength in JJA, the strongest PV gradients and the dynamical tropopause are forced into migrating to regions of the strongest winds, north of the split flow regime (see Bals-Elsholz et al. 2001). The reason for this follows from the fact that these tight gradients in PV must be in close proximity to the jet (Schwierz et al. 2004).

As a consequence, RWB events on the 330-K surface in the Southern Ocean are inhibited by the PFJ during DJF (Fig. 8a) and in the Australian–Pacific sector they are restricted by the STJ during JJA (Fig. 8c), as was the case at 350 K, but not to the same extent. The tropopause lies between the two jets during the intermediate seasons and so the jets do not inhibit the RWB in this latitude band. This results in the maximum in 330-K RWB counts during these seasons as discussed above.

Note also that most of the CWB events occur in the Australian–New Zealand–western Pacific Ocean sector during DJF and the equinox seasons, even though they are less than half as many as their anticyclonic counterparts (see Table 1). Peters and Waugh (2003) used barotropic shear arguments to show that double-jet structures that comprise both the STJ and PFJ caused the waves to break in an anticyclonic fashion, and the single-jet configurations that sometimes form when the STJ extends into the Pacific Ocean, without a PFJ component accompanying it to the south, lead to CWB. The formation of these wind configurations is highly variable.
on a submonthly scale. The occurrence of CWB is thought to be caused by single-jet structures that become apparent when 5-day mean basic states, for example, are considered as in Peters and Waugh (2003). This is the reason behind the increase in CWB from the 350- to the 330-K isentropic surface that was mentioned in section 3. On a seasonal time scale, the dominance of the double-jet structure is evident, hence the dominance of AWB in this sector.

In contrast to the cyclonic case, the AWB at 330 K has centers of occurrence at regions other than the sector encompassing the date line (i.e., South America, south of South Africa, and the southwestern coast of Australia; see Fig. 8). As before, this might have implications for the weather over these continental masses.

On the 320-K surface (not shown), AWB occurs mostly during winter, as noted in the previous section, and is concentrated southwest of Australia. It appears then that this RWB activity is influenced by the double jet, as pointed out in Peters and Waugh (2003). The summer maximum in 320-K CWB occurs on the poleward side of the polar front jet.

5. Summary and conclusions

A 30-yr climatology of Rossby wave breaking (RWB) on the Southern Hemisphere tropopause has been formed using NCEP–NCAR reanalysis fields. The mean structure and evolution of the RWB, as well as the seasonal and spatial variations in the occurrence of RWB, have been examined.

RWB events can be divided into four types, depending on whether they are anticyclonic or cyclonic and whether they are associated with equatorward or poleward fluxes of wave activity (Esler and Haynes 1999; GP08). Analysis of RWB in the Southern Hemisphere shows clear differences in the structure of PV and wave fluxes between anticyclonic (AWB) and cyclonic (CWB) events. The former have a northwest–southeast orientation, consistent with the anticyclonic southwest–northeast tilt shape of the Northern Hemisphere (e.g., Thornicroft et al. 1993; Lee and Feldstein 1996). The meridional wave activity fluxes associated with AWB are equatorward. In contrast, CWB events have a southwest–northeast tilt and the fluxes of wave activity are poleward. These differences in tilt are consistent with the direction of momentum fluxes, which is poleward and equatorward during AWB and CWB, respectively (Simmons and Hoskins 1980; Thornicroft et al. 1993; Rivière and Orlanski 2007).

Wave activity fluxes in AWB are slightly stronger than in CWB. This is expected since the eddy momentum fluxes are stronger in former than in the latter, as noted by Thornicroft et al. (1993).

However, the majority of RWB exhibits weak poleward–equatorward symmetry in terms of potential vorticity (PV), and the composite mean of the two types of AWB events (those that comply with LC1 and P2 conditions) are indistinguishable from a PV perspective (as are the two types of CWB events, LC2 and P1). This poleward–equatorward symmetry is consistent with discussion in Berrisford et al. (2007, see their Fig. 1). These predominant characteristics of RWB are consistent with the fact that the zonal flow is more zonally symmetric than in the NH. There are, however, differences in the meridional fluxes for LC1 and P2 events (and between LC2 and P1).

For example, the maximum LC1 fluxes are located on the equatorward side of the fold, where confluent flow can be expected, whereas the fluxes are located on the polar side for P2 events.

AWB events are much more common than CWB events on 350- to 330-K surfaces but are comparable on the 320-K surface. This occurs because the dynamical tropopause is generally within regions of anticyclonic shear, and as a result Rossby waves propagating along the tropopause tend to break in an anticyclonic fashion (Thornicroft et al. 1993; Peters and Waugh 1996, 2003). The position of the dynamical tropopause relative to the jet core is determined by the intersection of the PV = −1.5, −2.0, and −2.5 PVU surfaces with the isentropic surfaces, particularly the 330-, 340-, and 350-K surfaces.

The number of AWB events does not change much between surfaces, but the number of CWB events increases from the 350- to the 320-K isentropic surface. This increase in CWB events with lower isentropic surfaces was also observed by Martius et al. (2007).

There are large seasonal variations in the location and number of RWB events. On both surfaces and for both AWB and CWB events, there is a seasonal migration of the RWB activity from high latitudes during DJF to lower latitudes during JJA. This is closely linked to the migration of the dynamical tropopause (as seen also in the NH; e.g., Liniger and Davies 2004). The frequency of RWB also varies with season, but there is a sharp contrast in the seasonal variations on the two surfaces. On the 350-K surface there is a maximum count in DJF and a minimum in JJA, in agreement with Postel and Hitchman (1999), and this maximum extends to late March–May (MAM) and early September–November (SON) on the 340-K surface. In contrast, the maxima in RWB at 330 K occur during the equinox seasons, with the minimum occurring in DJF. On the 320-K surface AWB occurs most during the winter months and CWB is less variable during the year.

The migration of the dynamical tropopause and the related seasonal march of the tropospheric jets play a crucial role in influencing the seasonal occurrence of
RWB. The seasonal variation of 350-K RWB is caused primarily by changes in the frequency of RWB from the eastern Indian Ocean to western Pacific Ocean. During JJA the dynamical tropopause and subtropical jet are collocated in this sector and this inhibits RWB. In DJF the Rossby waves encounter no such obstruction and the maximum number of events occurs during this season. Away from the sector that is dominated by the subtropical jet, RWB has preferred regions of occurrence: the western Pacific Ocean, South Africa, and the Indian Ocean, with most of the events occurring over South Africa from autumn to spring. A large percentage of 350-K RWB events are anticyclonic and are characterized as such by the anticyclonic shear associated with the polar front jet, particularly those occurring over South Africa. The anticyclonic subtropical jet exit also plays a role also in inducing the breaking in the western Pacific Ocean from autumn to spring. The situation is very similar on the 340-K surface.

Variations in the jet locations also influence RWB at 330 K. The midlatitude dynamical tropopause and the polar front jet are collocated in the Southern Ocean during DJF, causing the minimum in RWB counts. The maxima during the equinox seasons are associated with the fact that the Rossby waves propagate and break between the polar front and subtropical jets. Even though the number of CWB events has increased on this surface as noted above, there are still about twice as many AWB events. Over South Africa the waves break anticyclonically because of the anticyclonic shear associated with the polar front jet during all seasons; however, in the other regions their anticyclonic sense is influenced by the double-jet structures (e.g., Peters and Waugh 2003) that prevail as a result of the development of the subtropical jet. However, on submonthly scales, single jets are established, causing the waves to break cyclonically in the Australian–eastern Pacific Ocean sector. This in turn results in the increased frequency of these types of RWB events from the 350- to the 330-K isentropic surface.

In this study we focused on the climatological structure and seasonal/spatial variability in RWB occurrence in the Southern Hemisphere. The interannual variations and trends of events will be examined in a future study.

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