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**Recommended Citation**


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Abstract To better understand the potential effects of climate change on atmospheric dynamics, this paper studies Rossby wave breaking and isentropic stratosphere-troposphere exchange (STE) in the Northern Hemisphere between 320 and 380 K during 1981–2015 using the Modern-Era Retrospective Analysis for Research and Application version 2 data. The isentropic STE is estimated using Contour Advection. Our results show that anticyclonic wave breaking events have become more frequent, especially in summer at higher isentropic surfaces, and cyclonic wave breaking events have become less frequent at 320 K. The anticyclonic wave breaking has shifted poleward in summer. The isentropic STE is found to be strongest over the regions where Rossby wave breaking activities are most frequent. Both isentropic STE and Rossby wave breaking are shown to be strongest in summer and weakest in winter. Our results do not show any discernable trends during isentropic STE during 1981–2015.

1. Introduction

One of the challenges in understanding the effects of climate change is that the dynamic response of the atmosphere to climate change is not well understood. Both modeling and observational studies have indicated a poleward shift of the subtropical jet stream resulting from the enhanced greenhouse effect (Archer & Caldeira, 2008; Davis & Rosenlof, 2012; Fu & Lin, 2011; Strong & Davis, 2007). This affects Rossby wave breaking (RWB), which is strongly related to the characteristics of the jet streams (Barnes & Hartmann, 2012; Horinouchi et al., 2000).

RWB is an important dynamical process associated with weather systems such as cutoff lows and blocking anticyclones (Baray et al., 2003; Ndarana & Waugh, 2010; Pelly & Hoskins, 2003), storm track (Robinson, 2006), and North Atlantic weather regimes (Michel & Rivièrè, 2011; Woollings et al., 2008). During an RWB event, potential vorticity (PV) contours become severely distorted, and sections of the contours stretch away from the dynamical barriers (i.e., the edges of the polar vortex and the tropopause), extending equatorward into the tropical latitudes or poleward into higher latitudes (Appenzeller et al., 1996; McIntyre & Palmer, 1983; Norton, 1994). In this way, RWB can induce stratosphere-troposphere exchange (STE) processes (Holton et al., 1995; Norton, 1994; Postel & Hitchman, 1999). The STE, in turn, can result in the exchange of momentum, heat, and chemical species between the upper troposphere and lower stratosphere (UTLS). Several studies have shown that RWB is associated with the isentropic transport of air masses and trace gases such as ozone (Appenzeller et al., 1996; Jing et al., 2004; Liu & Barnes, 2018; Scott & Cammas, 2002). Therefore, understanding the occurrence of RWB in the UTLS region is important to the study of atmospheric chemistry and the radiative energy budget, as well as atmospheric dynamics. It will also help to predict how weather events will change under a changing climate.

RWB events are classified into two types: anticyclonic wave breaking (AWB) and cyclonic wave breaking (CW). AWB events are typically associated with equatorward transport and CWB with poleward transport (Magnúsdóttir & Haynes, 1996; Thornicroft et al., 1993). The occurrence of RWB is affected by the jet stream. AWB typically occurs equatorward of the jet and CWB poleward of the jet (Weijenborg et al., 2012). When the jet stream shifts poleward, the frequency of AWB increases and the frequency of CWB decreases (Liu & Barnes, 2018; Rivièrè, 2011; Rivièrè et al., 2010). This may also change the latitudinal position of RWB events. This work has two major objectives: (1) to study how the occurrences of the two types of RWB have changed since 1981 in terms of their frequency and position and (2) to analyze the relationship between RWB and isentropic STE over the same period. This study focuses on isentropic surfaces between 320 and 380 K, which are...
located within the so-called “middle world,” where the isentropic surfaces intersect the tropopause and isentropic transport often results in the exchange of air and chemicals between the tropical upper troposphere and the extratropical lower stratosphere (Holton et al., 1995). Many previous studies have focused on single isentropic layers (e.g., Postel & Hitchman, 1999; Strong & Magnusdottir, 2008; Waugh & Polvani, 2000), shorter periods (e.g., Peters & Waugh, 2003; Postel & Hitchman, 1999), or cold seasons (e.g., Martius et al., 2007; Michel & Rivièreme, 2011; Peters & Waugh, 2003; Strong & Magnusdottir, 2008; Woollings et al., 2008). In this work, we study RWB on multiple isentropic layers over an extended period of time and in different seasons.

RWB has been studied by numerical modeling experiments (e.g., Barnes & Hartmann, 2012; Liu & Barnes, 2018; Norton, 1994; Rivièreme, 2009; Waugh & Plumb, 1994) and by climatological analysis of observation-based reanalysis data sets (e.g., Hitchman & Huesmann, 2007; Martius et al., 2007; Postel & Hitchman, 1999). The data set used most often in previous work is the 40-year European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40). The ERA-40 products have been used to study the climatology of wintertime RWB in the Northern Hemisphere below 350 K (Martius et al., 2007) and the association between RWB and wintertime weather regimes (Michel & Rivièreme, 2011; Woollings et al., 2008). Other studies have employed National Centers for Environmental Prediction and the United Kingdom Met Office reanalysis fields to study the seasonal climatology of RWB between 320 and 2,000 K in 1979–2005 (Hitchman & Huesmann, 2007) and the association of RWB with wintertime North Atlantic Oscillation and Northern Hemispheric Anular Mode (Strong & Magnusdottir, 2008). Homeyer and Bowman (2013) studied the isentropic transport between the tropics and extratropics associated with RWB in the 350–500-K potential temperature range (which is above the subtropical jet) in 1981–2010 using the European Centre for Medium-Range Weather Forecasts Interim Re-Analysis data. Because of the important roles that RWB plays in the weather regimes and the redistributions of heat and trace gases, we need to fully understand the occurrence of RWB and its association with isentropic STE.

We use the meteorological fields during 1981–2015 from the Modern Era Retrospective Analysis for Research and Applications version 2 (MERRA-2) reanalysis to study both RWB and isentropic STE in the 320–380-K potential temperature range. The dynamic characteristics of the UTLS, including the subtropical jet streams and the tropopause, are well represented in MERRA products (Manney et al., 2014). This study analyzes the 35-year trend of RWB (both AWB and CWB) and isentropic STE. Our results (1) help to understand the effect of climate change on RWB, (2) demonstrate how well RWB is represented in MERRA-2, and (3) provide greater confidence about the relationship between RWB and isentropic STE.

2. Reanalysis Data

This study uses meteorological fields for the period 1981–2015 from the Modern-Era Retrospective Analysis for Research and Applications version 2 (MERRA-2) reanalysis obtained from the Goddard Earth Sciences Data and Information Services Center. MERRA-2 data are available on 42 pressure levels from 1,000 to 0.1 hPa at the resolution of 0.5° (latitude) × 0.625° (longitude) every 3 hr from 00 GMT to 21 GMT. The spatial and temporal resolutions of MERRA-2 data are adequate to address RWB events, which typically occur at large scales (>1,000 km) over 24 hr. Through intercomparison with other reanalysis data sets, including the European Centre for Medium-Range Weather Forecasts Interim Re-Analysis and the Climate Forecast System Reanalysis, it has been shown that MERRA products have good correlation with other data sets (Rienecker et al., 2011). The spatial and temporal frequency distributions of upper tropospheric jets from MERRA data are consistent with the results of other studies using different data sets; jet stream dynamics and the PV-defined tropopause have been shown to be accurately represented in MERRA (Manney et al., 2014).

This study focuses on the upper troposphere and the lower stratosphere (UTLS), which we define as the region located between 320- and 380-K isentropic surfaces, which intersect PV surfaces at midlatitudes (Figure 1). As described in section 3.1, PV surfaces are used to define the tropopause. Isentropic transport on surfaces between 320 and 380 K often results in cross-tropopause exchange of air and chemical species. We do not include isentropic surfaces below 320 K, because these surfaces intersect the tropopause mainly in wintertime and rarely in summertime (Figure 1); including these surfaces would add few results for understanding the seasonality and long-term trend of isentropic STE. The MERRA-2 data set has been shown to have good accuracy in representing the UTLS dynamic fields (Gelaro et al., 2017; Manney et al., 2014). In order to study the RWB and isentropic STE, the MERRA-2 data (including PV and winds) from 1981 to 2015 are...
interpolated vertically from pressure surfaces onto isentropic surfaces (from 320 to 380 K at 10-K vertical resolution) using the interpolation method described in Edouard et al. (1997). We then focus on the three layers centered at 330, 350, and 370 K.

3. Methodology

3.1. Definition of the Tropopause

The dynamical tropopause marks the boundary between the troposphere and the stratosphere where isentropic transport is hindered. It is often defined by a PV value on isentropic surfaces where the strongest latitudinal gradient of PV is found—otherwise, the strong PV gradient would be destroyed by vigorous mixing of the air (Holton et al., 1995). The PV values used to define the dynamical tropopause range between 1.5 and 5.0 PVU (1 PVU = 1.0 × 10⁻⁶ m² s⁻¹ K⁻¹ kg⁻¹; Highwood et al., 2000; Hoerling et al., 1991; Manney et al., 2011; Postel & Hitchman, 1999; Schoeberl, 2004). The PV value that marks the greatest gradient in PV on isentropic surfaces increases with increasing potential temperature (Homeyer & Bowman, 2013; Kunz et al., 2011). The boundary between the tropics and extratropics at 350 K is best represented by 4.0-PVU contours (Homeyer & Bowman, 2013). This study uses these PV values (3, 3, 4, 4, 5, 5, 6 PVU) to define the tropopause on 320, 330, 340, 350, 360, 370, and 380 K, respectively.

3.2. Identification of RWB

Rossby waves occur because of the variation of the Coriolis force with latitude and the conservation of absolute vorticity (Holton, 1992). RWB occurs when the amplitude of the wave reaches a critical point at which the crest or trough of the wave overturns. RWB is characterized by the rapid and irreversible deformation of material contours and is often identified by the reversed latitudinal gradient of PV (Postel & Hitchman, 1999). Based on the wave-breaking detection algorithms in Strong and Magnusdottir (2008) and Homeyer and Bowman (2013), by analyzing MERRA-2 PV fields on isentropic surfaces, RWB is detected and identified as anticyclonic or cyclonic wave breaking as described below (shown in Figure 2). We search for RWB events on the tropopause, which is defined as the longest closed PV contour that circles around the North Pole on an isentropic surface. The PV values of the tropopause (PV*) are defined as described in the previous section.

First, on each isentropic surface, we search for the sections on the PV* contour that experienced RWB. PV contours are quasi-parallel to the latitude circles, and the absolute value of PV increases from the equator to the pole for a standing Rossby wave. When RWB occurs, PV contours will overturn and the local latitudinal PV gradient will change sign accordingly. As shown in Figure 2, the overturning of a PV contour is where the meridian intersects the PV contour at least 3 times. We record the time and positions (latitude and longitude) of the nodes on the overturning sectors of the PV* contour. Overturning points within 1,000 km are considered to be part of the same RWB event.

Next, we distinguish between anticyclonic and cyclonic RWB events. During a RWB event, as the air flows eastward, the PV contours can wrap in either anticyclonic or cyclonic directions. An AWB event is characterized by the wave tilting in the northeast-to-southwest direction. The westernmost overturning point (PW) on the PV contour is located poleward of the easternmost overturning point (PE). On the other hand, during CWB, the PV contours wrap cyclonically, tilting in the southeast-to-northwest direction, and PW is equatorward of PE (Figure 2). For each AWB and CWB event, based on the times and positions of the nodes on the overturning sectors of the PV* contour, we calculate the location of its centroid, its duration, and its longitudinal and latitudinal extent. Only irreversible AWB and CWB events are considered for the RWB analysis, in order to avoid including small patches of short-lived PV reversals that are not part of a major RWB event. Criteria for irreversible RWB are the following: longitude extent >10°, latitude extent >5°, and duration ≥24 hr (Homeyer & Bowman, 2013).

We then analyze the frequency, geographical location, and seasonal variation of both AWB and CWB on each isentropic surface in 320–380 K during 1981–2015.
3.3. Quantification of Isentropic STE

We investigate the isentropic STE during 1981–2015 using the trajectory model called Contour Advection with Surgery (CAS) that was developed by Waugh and Plumb (1994). Contour Advection is able to capture the continuous formation of small-scale structures in a passive tracer field (e.g., a PV field) on isentropic surfaces with a spatial resolution much finer than that of the analyzed data (Baker & Cunnold, 2001; Waugh & Plumb, 1994). Its surgery routine removes the fine-scale structures smaller than a prescribed cutoff scale. These removed small-scale structures are assumed to have been irreversibly mixed with the air into which they are transported. The CAS model has been used to quantify the isentropic exchange of air mass, water vapor, and O3 (Dethof, O'Neill, Slingo, 2000; Dethof, O'Neill, Slingo, Berrisford, 2000; Jing et al., 2004, 2005).

Following the quantification method in Jing et al. (2004), we next calculate the fluxes of isentropic STE on each of the isentropic surfaces (330, 350, and 370 K). We select the five PV contours (2, 3, 4, 5, 6 PVU) to run CAS, because 90% of the isentropic STE are associated with PV values between 2 and 6 PVU (Jing et al., 2004). First, we find the initial locations (the longitude and latitude) of the longest pole-circling contours of the five PV values on the isentropic surface from the MERRA-2 PV fields on day 0. We run the model without surgery for five days, with the MERRA-2 wind fields being renewed every 6 hr. The PV field on day 5 generated by Contour Advection shows the presence of filaments, demonstrating the small-scale structure associated with the isentropic transport. We also run the surgery routine of CAS to remove the small-scale structures and thus find the position of the tropopause on day 5. The PV field with small-scale structures on day 5 is then compared with the position of the tropopause on day 5. The small-scale structures are identified as stratosphere-to-troposphere (S-T) transport if they are located on the equator side of the tropopause and their PV values are greater than the PV value of the tropopause (PV*). Conversely, those on the poleward side of the tropopause with PV < PV* are identified as troposphere-to-stratosphere (T-S) transport. The daily isentropic S-T (T-S) fluxes (in kg/day) are estimated on a 0.25° × 0.25° grid by calculating the masses of these small-scale structures divided by five days. The daily net flux is the difference between the ST and TS fluxes.

3.4. The Mann-Kendall Trend Test

The Mann-Kendall trend test (Kendall & Stuart, 1967; Mann, 1945) is a nonparametric test and is commonly used to detect monotonic trends in series of environmental data and climate data (Crawford et al., 1983; Hirsch et al., 1982; Steele et al., 1974; Yue et al., 2002). Since the test is a nonparametric one, it does not assume any joint distribution for the data and is not affected by departure from normality. The null hypothesis ($H_0$) for this test states that the observations are simply independent and identically distributed. $H_0$ is tested against an alternative hypothesis ($H_1$) that there is a monotonic trend in the data (one-sided alternative may be defined to test positive or negative trend). For $n$ observations, we have $n(n - 1)/2$ pairs $(X_i, X_j)$ and the MK test statistic is defined as
\[ S = \sum_{i} \sum_{j \neq i} \text{sign}(X_j - X_i) \]

where \( \text{sign}(X_j - X_i) \) can be +1, −1, or 0 when \( X_j >, <, \) or \( =X_i \), respectively. \( S \) has mean 0 and variance

\[ \sigma^2 = \left( \frac{n(n-1)(2n+5) - \sum_{j=1}^{T} (t_j - 1)(2t_j + 5)}{18} \right) \]

where \( T \) is the number of tied groups and \( t_j \) is the number of observations in the \( j^{th} \) group. It may be shown that \( S \) is approximately normally distributed. In this paper, all tests were performed for a level of significance of 5%. When testing for positive or negative trend, we define our alternative hypothesis \( H_1 \) accordingly and reject \( H_0 \) to conclude that there is a positive/negative trend if the \( p \) value of the test is less than \( \alpha \) (\( \alpha = 0.05 \)).

**4. Results of Rossby Wave Breaking Analysis**

**4.1. Frequency of RWB**

Using the method described in section 3.2, we have identified irreversible RWB events in the Northern Hemisphere on each isentropic surface in the 320–380-K range during 1981–2015 and distinguished between AWB and CWB events. The number of AWB events is 1 order of magnitude greater than that of CWB on all these isentropic surfaces (see the results for 320 and 350 K in Figure 3, for example). This is consistent with previous studies, which have shown that about 90% of RWB events are anticyclonic (Homeyer & Bowman, 2013; Rivière, 2009). From 1981 to 2015, we find that the annual number of AWB events increased by 0.74 ± 0.19 events per year \( (p < 0.05) \) at 350 K but decreased by 0.54 ± 0.20 events per year \( (p < 0.05) \) at 320 K; such trends are found in both summer and winter (Figure 3). The linear fit of the AWB events with year yields positive slopes at all isentropic levels except 320 K and the change is most steep at 350 K (Figure 4). The positive trends are statistically significant \( (p < 0.05) \) for annual AWB events at 350 K and summer AWB events.
at 370 and 380 K; the negative trend of AWB at 320 K is statistically significant for both annual and summer events (Table 1). Similar to the AWB results, our analysis of CWB events also shows positive trends except at 320 K, where CWB events decreased during 1981–2015 (Figure 4).

Generally, we find that there is an increasing trend of both AWB and CWB from 1981 to 2015 above 320 K and a decreasing trend at 320 K; such trends are more statistically significant in summer than in winter. The increase of AWB events above 320 K may be the result of strengthening of the Brewer-Dobson circulation, a major mechanism that drives the meridional transport in the stratosphere. The Brewer-Dobson circulation has been intensified by the enhanced greenhouse effect, resulting in stronger upwelling in the tropical UT and stronger subsidence in the extratropical LS (Butchart et al., 2006). The strengthened tropical convection in a warmer troposphere causes an increased equatorward drag for the AWB to occur in the extratropical LS, and this mechanism is most effective in summer months (Deckert & Dameris, 2008; Shepherd & McLandress, 2011). Furthermore, the increased AWB drag, in turn, can lead to stronger Brewer-Dobson circulation (Shepherd & McLandress, 2011). Therefore, AWB becomes more frequent in the LS especially in summer, as our results demonstrate. We also note that the change in the AWB frequency is most steep at 350 K (Figure 5). This is likely due to the fact that the 350-K surface is close to the core of the subtropical jet (Homeyer & Bowman, 2013; Michel & Rivière, 2011), and RWB has been shown to be associated with the characteristics of the jet (Barnes & Hartmann, 2012). Therefore, RWB may be most sensitive to the changes in the jet stream at 350 K.

![Figure 4](image-url). The linear trends of (left) AWB and (right) CWB events during 1981–2015 at isentropic surfaces between 320 and 380 K.

<table>
<thead>
<tr>
<th>Potential temperature (K)</th>
<th>AWB events</th>
<th>CWB events</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Annual</td>
<td>Summer</td>
</tr>
<tr>
<td>380</td>
<td>0.275</td>
<td><strong>0.009</strong></td>
</tr>
<tr>
<td>370</td>
<td>0.193</td>
<td><strong>0.041</strong></td>
</tr>
<tr>
<td>360</td>
<td>0.239</td>
<td>0.706</td>
</tr>
<tr>
<td>350</td>
<td><strong>0.000</strong></td>
<td>0.066</td>
</tr>
<tr>
<td>340</td>
<td>0.319</td>
<td>0.133</td>
</tr>
<tr>
<td>330</td>
<td>0.523</td>
<td>0.811</td>
</tr>
<tr>
<td>320</td>
<td>0.993</td>
<td>0.998</td>
</tr>
</tbody>
</table>

Table 1
The p Values for the Mann-Kendall Trend Test on the Frequencies of AWB and CWB Events During 1981–2015

Note. The tested hypotheses are the following: $H_0$: there is no trend versus $H_a$: positive (increasing) trend. The p values that are smaller than 0.05 are bolded.
Previous studies have shown that the frequency of RWB is closely linked to the latitudinal fluctuation of the jet streams (Rivière, 2011; Rivière et al., 2010). The enhanced upper tropospheric baroclinicity due to global warming can result in a poleward shift of the jet streams (Rivière, 2011). Fu and Lin (2011) showed that the subtropical jets have shifted poleward by 1° latitude in the Northern Hemisphere since 1979. Modeling studies indicate that the higher the jet latitude is, the more frequent the AWB is and the less frequent the CWB is (Liu & Barnes, 2018; Rivière, 2011; Rivière et al., 2010). Our results here show the observed trends of RWB in MERRA-2, which are not totally consistent with the modeling results. Our results are consistent with them in the way that AWB occurrence increases in the levels above the subtropical jet (which is at ~350 K) and CWB occurrence decreases in the levels below the jet (i.e., at 320 K). However, our results also show an increase in CWB frequency above 320 K. Previous studies on the relationship between RWB and the jet stream mainly focus on the layers below 350 K (Liu & Barnes, 2018; Rivière, 2011; Rivière et al., 2010). Our results provide information on the observed changes in RWB occurrence in 320–380 K. The sensitivity analysis (provided in section 4.4) shows that using different PV tropopause values will not change the general conclusion on the trends of RWB occurrence.

4.2. Geographical Distribution of RWB

Based on the locations of the centroid of RWB events, we calculated the monthly mean frequencies of AWB and CWB during 1981–2015 on a 2.5° × 2.5° grid at each isentropic level. Figure 5 shows the geographical distribution of AWB and CWB during summer (JJA) and winter (DJF) at 330, 350, and 370 K.

Figure 5. The mean frequency of anticyclonic wave breaking in summer (JJA) and winter (DJF) during 1981–2015 at 330, 350, and 370 K.
distributions of AWB occurrence in summer (June-July-August) and winter (December-January-February) at the three isentropic levels: 330, 350, and 370 K. Similar results for CWB are shown in Figure 6. In general, the frequencies of both AWB and CWB are higher in summer than in winter. This is because the strong latitudinal PV gradients around the tropopause tend to inhibit RWB, and such barriers are weaker in summer than in winter (Hitchman & Huesmann, 2007). Therefore, RWB (including both AWB and CWB) tends to occur more frequently in summer.

The AWB in winter occurs most frequently over the eastern North Atlantic and the eastern North Pacific. This agrees with the wintertime wave breaking that was derived from PV streams using the ERA-40 data set (Martius et al., 2007). In summer, high AWB frequencies are found over the western North Pacific and eastern Asia, although the frequency maxima are mainly found in the region extending from the eastern North Pacific, across North America, and into the North Atlantic. Our results are consistent with Scott and Cammas (2002), which shows that the strongest RWB in the Northern Hemisphere occurs over the eastern Atlantic and northwest Africa in winter and over the western Pacific in summer. The RWB-induced tropopause folding at 350 K also tends to occur over the summertime oceans and downstream of the subtropical high-pressure systems (Postel & Hitchman, 1999). Our results also show that, in general, AWB (CWB) frequencies increase (decrease) with increasing potential temperature (Figures 5a and 5b), which is consistent with the findings of Martius et al. (2007) and Rivière (2009). This is because RWB is controlled by two competing

Figure 6. Same as in Figure 5 but for cyclonic wave breaking.
factors: the absolute vorticity term and the stretching term. AWB is affected more by the absolute vorticity term, which is more efficient at upper levels, and CWB is affected more by the stretching term, which is more efficient at lower levels (Rivière, 2009).

Summer AWB frequencies are shown to increase with increasing potential temperature; winter AWB frequencies, on the other hand, are stronger at lower isentropic surfaces (Figure 5). This is consistent with the results of Homeyer and Bowman (2013), who studied RWB in the 350–500-K range during 1981–2010 using the European Centre for Medium-Range Weather Forecasts Interim Re-Analysis data set. According to Homeyer and Bowman (2013), the number of RWB events started to increase with increasing potential temperature above 360 K and RWB generally peaked in summer to fall. At lower isentropic levels (e.g., 330 K), the AWB occurs more frequently in winter. This may be because AWB is controlled more by the absolute vorticity term (Rivière, 2009) which is stronger at lower levels in winter. It could also be because RWB activities ascend well into the stratosphere in summer (Homeyer & Bowman, 2013), but not so well in winter due to stronger dynamical barriers of the tropopause and polar vortex.

4.3. Latitudinal Migration of RWB Events

For each AWB and CWB event, we track the location of its centroid. We then calculate the mean latitude of the centroids of these events. Figure 7 shows that AWB events at 350 K have shifted poleward in summer (June-July-August) and equatorward in winter (December-January-February). Such shifts of AWB are found at most isentropic levels (as shown in Figure 8, in which positive slopes indicate poleward shifts and negative slopes indicate equatorward shifts). The results for CWB are similar to those for AWB (Figure 8), demonstrating a poleward shift in summer and an equatorward shift in winter. We also find that at 320 K both AWB and CWB in both summer and winter show a poleward shift. As shown in the previous section, the frequency of both AWB and CWB decreased in both summer and winter. This is because CWB mainly occurs on the poleward flank of the subtropical jet stream. A more northward jet makes CWB less probable (Michel & Rivière, 2011; Rivière, 2009).

Figure 7. The mean latitude of the center of the AWB events on 350 K in summer and winter. The identification of the RWB is based on the 4-PVU tropopause definition.

Figure 8. The linear trends of the mean latitude for (left) AWB and (right) CWB during 1981–2015 at isentropic surfaces between 320 and 380 K.
We then conducted the Mann-Kendall trend tests on the mean latitudes of AWB and CWB events. The tested hypotheses for summer are the following: \( H_0: \) there is no trend versus \( H_1: \) positive (poleward) trend, and for winter: \( H_0: \) there is no trend versus \( H_1: \) negative (equatorward) trend. We reject \( H_0 \) and conclude \( H_1 \) that there is a positive or negative trend if the \( p \) value of the test is less than 0.05. The results are listed in Table 2. The summer poleward shift of AWB is significant at most isentropic surfaces, and the poleward shift of CWB is significant only at potential temperatures below 350 K. The winter equatorward shift is significant only at 370 K for AWB and at 340 and 360 K for CWB. Therefore, the equatorward shift of the winter wave breaking events is much weaker than the poleward shift of summer events.

Overall, for AWB the summer events have a significant poleward shift for most temperatures, while for CWB the same is true for the lower temperatures only. In winter, a significant equatorward shift of frequencies is seen only for one or two temperatures, for both AWB and CWB. This is consistent with the results for the Southern Hemisphere (Barnes & Hartmann, 2012), which showed that as the jet shifts poleward both AWB and CWB shift poleward but CWB shifts less significantly. Fu and Lin (2011) showed that the subtropical jets have shifted poleward by 1.0° ± 0.3° latitude in the Northern Hemisphere since 1979, which is consistent with our findings.

### 4.4. Sensitivity of Identified RWB to the Tropopause Definition

In order to test how sensitive our results are to the tropopause definition, we compare the results of different PV values used to define the tropopause at 350 K (Tables 3 and 4). Both AWB and CWB frequencies increase with increasing PV value; the average latitudes of both AWB and CWB also increase with increasing PV value (Table 3). Although the slopes of the trends (both the number of events and the latitudinal shifts) vary in values for different PV tropopause values, the general direction of the trends stays the same (Table 4), indicating that the direction of the change is not sensitive to the tropopause definition.

### 5. Isentropic Transport Across the Tropopause

Using the quantification method based on CAS described in section 3.3, we calculated the fluxes of cross-tropopause transport in both S-T and T-S directions on 330, 350, and 370 K. Figure 9 shows the annual results as well as the monthly average fluxes. The estimated annual S-T fluxes are always greater than the T-S fluxes, although they are on the same order of magnitude, resulting in net S-T transport at all three isentropic surfaces during 1981–2015. This means that isentropic STE is a two-way process, but its net effect is to transfer air from the extratropical LS into the subtropical UT quasi-horizontally. The isentropic transport of air is stronger in the direction from S-T than in the direction from T-S in the midlatitudes (Dethof, O’Neill, Slingo, 2000; Sprenger & Wernli, 2003; Yang et al., 2016). Our results for STE during 1981–2015 extend the findings of Jing et al. (2004, 2005), which looked at the isentropic exchange of ozone in individual years 1990 and 1999. The results of this study confirm that the isentropic STE occurs primarily where RWB occurs, indicating that RWB is a primary driver for isentropic transport.
Both the S-T and T-S monthly fluxes are strongest in summer (June-July-August; Figure 9), which is similar to the seasonality of RWB occurrence. The magnitude of the net isentropic transport also peaks in summer, and its direction is S-T in every month except April. This seasonality of isentropic STE is opposite to that of the downward diabatic STE at midlatitudes, which is strongest in summer and weakest in winter (Olsen et al., 2002, 2003). The diabatic STE is part of the Brewer-Dobson circulation, which is strongest in winter (Holton et al., 1995). On the other hand, the isentropic mixing across the tropopause is strongest in summer because the extratropical PV gradients are smallest in summer and thus act as much weaker barriers to the isentropic STE (Chen, 1995; Haynes & Shuckburgh, 2000).

Figure 9 also shows that the fluxes of isentropic transport in both directions are largest at 330 K and decrease with increasing altitude. As shown in section 4, the frequencies of RWB, on the other hand, increase with increasing potential temperature. There are several reasons for this. First, the isentropic density of the air decreases more rapidly with increasing altitude than the increase of RWB frequency. Second, the tropopause barrier is not as strong on the lower surfaces as on the upper ones, and this results in greater isentropic mixing.

Table 4
The Linear Trends of the Frequencies and the Latitudinal Shifts of AWB and CWB During 1981–2015 at 350 K for Different PV Tropopause Values

<table>
<thead>
<tr>
<th>PVU</th>
<th>AWB events (events/year)</th>
<th>AWB latitude (deg/year)</th>
<th>CWB events (events/year)</th>
<th>CWB latitude (deg/year)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Annual</td>
<td>JJA</td>
<td>DJF</td>
<td>Annual</td>
</tr>
<tr>
<td>3</td>
<td>0.45 (0.17)</td>
<td>0.23 (0.06)</td>
<td>0.51 (0.08)</td>
<td>0.01 (0.02)</td>
</tr>
<tr>
<td>4</td>
<td>0.74 (0.19)</td>
<td>0.12 (0.09)</td>
<td>0.15 (0.13)</td>
<td>0.02 (0.02)</td>
</tr>
<tr>
<td>5</td>
<td>0.67 (0.26)</td>
<td>0.18 (0.09)</td>
<td>0.02 (0.13)</td>
<td>0.01 (0.02)</td>
</tr>
</tbody>
</table>

Note. The standard deviations of the slopes are in parenthesis. The trends that are statistically significant (p value <0.05) are bolded.

Figure 9. Estimated (left column) annual total and (right column) mean monthly isentropic fluxes during 1981–2015 for (bottom row) 330 K, (middle row) 350 K, and (top row) 370 K. Positive (negative) net fluxes indicate net S-T (T-S) transport.
mixing on the lower surfaces (Chen, 1995; Dethof, O’Neill, Slingo, 2000). Third, the mixing strength per RWB event may decrease with increasing isentropic surfaces.

Unlike the RWB occurrence, our estimated fluxes of isentropic transport do not show any significant trends during 1981–2015. This suggests that although isentropic STE is induced by RWB, as indicated by their seasonality and geographic distributions, the amount of cross-tropopause transport is also influenced by other factors, such as the characteristics of the tropopause. Theoretically, more RWB events should increase the probability of isentropic STE. However, our quantification of isentropic STE is also dependent on the tropopause. In section 4.4, we have shown that the frequencies of both AWB and CWB occurrence increase with increasing tropopause PV value. The estimated isentropic fluxes, on the other hand, decrease with increasing PV value (Jing et al., 2004), because the areas of the small-scale structures that are identified as cross-tropopause transport decrease at higher PV values (Scott & Cammas, 2002). This means that the increasing effect on the STE flux caused by more frequent RWB could have been offset by the decreasing effect caused by higher PV tropopause values. More research is needed to further investigate if the PV tropopause has also moved poleward and consequently has a greater PV value since 1981.

Using the gridded daily isentropic fluxes that are quantified as described in section 3.3, we calculate the seasonal mean S-T (T-S) isentropic fluxes during 1981–2015. Figure 10 shows the geographical distributions of the average isentropic S-T and T-S transport at 350 K in summer and winter during the period 1981–2015. The S-T transport is stronger than T-S, and both S-T and T-S are stronger in summer than in winter. In winter, the S-T transport occurs predominantly in the midlatitude regions of the eastern North Atlantic and northwest Africa.

Figure 10. Geographical distributions of the estimated average isentropic cross-tropopause transport during 1981–2015 for 350 K in the (top row) S-T and (bottom row) T-S directions in summer (JJA) and winter (DJF).
In summer, the S-T fluxes are located primarily over the North Pacific Ocean and have secondary maxima over the North Atlantic and the southeast United States, where RWB preferentially occurs around 350 K in summer (as previously shown in Figure 5). Although smaller in magnitude, the T-S fluxes are shown to occur mainly over the Pacific Ocean and the Mediterranean Sea in summer and over the Atlantic in winter.

6. Conclusions

In this paper, we have studied the two types of Rossby wave breaking (AWB and CWB) and their association with isentropic STE in the upper troposphere and lower stratosphere (UTLS) region using the MERRA-2 data set over the 35-year period between 1981 and 2015. This is one of the longest periods for which these phenomena have been studied. By studying the frequency, seasonality, and geographical locations of both AWB and CWB during 1981–2015, our results provide conclusions on both types of RWB over an extended period on several isentropic surfaces (320–380 K) in both summer and winter. We have also estimated the isentropic transport and studied its association with RWB over the same 35-year period.

Our major findings on RWB are as follows: (a) both AWB and CWB have become more frequent since 1981 above 320 K and less frequent at 320 K, (b) the mean latitude of AWB and CWB events has shifted poleward, and (c) such changes in frequency and latitude are more significant in summer than in winter. Our results on isentropic STE show (a) the isentropic S-T transport is stronger than the T-S transport, resulting in net S-T transport at all isentropic surfaces; (b) both S-T and T-S transport are strongest in summer and weakest in winter; and (c) the strongest S-T transport occurs over the North Atlantic and North Pacific Oceans in summer and over the North Atlantic in winter. We also find that isentropic STE and RWB have the same seasonality and occur in similar locations. However, our results do not show discernable trends in the magnitude of isentropic STE during 1981–2015.

This work confirms that MERRA-2 reliably captures the dynamical characteristics of the atmosphere in the UTLS region. Our results for RWB are also in agreement with previous studies related to the poleward shift of the jet streams. Our results on the change of RWB in the 1981–2015 period can help to assess the potential impact of climate change on the dynamics of the atmosphere. However, there are some caveats that could be addressed in future work. First, the isentropic surfaces could be expanded beyond 320–380 K, which would help us understand how RWB may have changed outside the tropopause region. Second, other reanalysis data products could be used to do similar analyses, which may enable the period to be extended and provide greater confidence in the long-term changes of RWB and STE. Additionally, because of the important roles that RWB plays in STE, it is important to understand how RWB-induced STE will affect the exchange of chemicals such as ozone. In future work, we plan to study the impact of RWB and isentropic STE on tropospheric ozone levels.

Acknowledgments

P. Jing is supported by funding from the National Aeronautics and Space Administration under grant NNX16AG36G. We acknowledge NASA’s Goddard Global Modeling and Assimilation Office for producing the MERRA-2 products. The MERRA-2 fields used in this study are provided at https://goldsmr5.gesdisc.eosdis.nasa.gov/data/MERRA2/M2I3NPASM.5.12.4. The derived data products are available at https://www.luc.edu/sustainability/about/staff/rossbywavedata/. We also thank the two anonymous reviewers for their constructive comments to improve the manuscript. There are no real or perceived financial conflicts of interests for either author.

References


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