Synoptic formation of double tropopauses

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Key Points:

\begin{itemize}
\item Double tropopauses are closely linked to Rossby wave breaking.
\item Double tropopauses can form as a result of differential advection and/or the destabilization of thermal stratification.
\item The occurrence of double tropopauses substantially enhances stratosphere-troposphere exchange of air mass.
\end{itemize}

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Abstract

Double tropopauses are ubiquitous in the midlatitude winter hemisphere and represent the vertical stacking of two stable tropopause layers separated by a less stable layer. By analyzing COSMIC GPS data, reanalysis, and eddy lifecycle simulations, we demonstrate that they often occur during Rossby wave breaking and act to increase the stratosphere-to-troposphere exchange of mass. We further investigate the adiabatic formation of double tropopauses and propose two mechanisms by which they can occur. The first mechanism operates at the tropopause break in the subtropics where the higher tropical tropopause sits on one side of the break and the lower extratropical tropopause sits on the other. The double tropopauses are then formed by differential meridional advection of the higher and lower tropopauses on the two sides of the tropopause break. We show that anticyclonic wave breaking can form double tropopauses mainly by providing stronger poleward advection of the higher tropopause in its poleward lobe. Cyclonic wave breaking mainly forms double tropopauses by providing stronger equatorward advection of the lower tropopause in its equatorward lobe. We demonstrate in the COSMIC GPS data and reanalysis that about half of the double tropopauses in the northern hemisphere winter can be directly attributed to such differential advection. For the second mechanism, adiabatic destabilization of the air above the tropopause contributes to the formation of a double tropopause. In this case, a tropopause inversion layer (TIL) is necessary for this destabilization to result in a double tropopause.

1 Introduction

The tropopause marks an interface between two dynamically and chemically distinct parts of the atmosphere – the stratosphere and the troposphere. The transport and mixing of compositions across this interface has profound consequences for the global climate and chemistry. For example, the variation in greenhouse gases such as water vapor and ozone near the tropopause is strikingly efficient in altering the radiative forcing and hence the global surface temperature [e.g. Forster et al., 1997; Solomon et al., 2010; Riese et al., 2012]. Furthermore, the variability of stratosphere-to-troposphere ozone flux can translate to the variability of the health-related ultraviolet index [e.g. Hegglin and Shepherd, 2009]. In addition to its radiative impact, ozone is itself detrimental to human health and its surface concentration is currently regulated by the U.S. Environmental Protection Agency (U.S.)
The structure of the tropopause is thus of great interest due to its ability to influence the exchange between the stratosphere and the troposphere. Particularly, the vertical folding of the tropopause is an important candidate for enhancing this exchange [e.g. Shapiro, 1980]. When observed via sounding profiles, this folding structure appears as “double tropopauses” [e.g. Kochanski, 1955; Seidel and Randel, 2006], featuring the vertical stacking of two stable tropopause layers separated by a less stable layer (Fig. 1a).

Double tropopauses occur in both hemispheres and all seasons, with their highest frequencies in winter [e.g. Randel et al., 2007a; Añel et al., 2008; Peevey et al., 2012]. During double tropopause events, radiosonde and satellite observations above the first tropopause show less stratospheric trace gases and more tropospheric trace gases compared to single tropopause events [e.g. Randel et al., 2007a; Schwartz et al., 2015], indicating strengthened transport and mixing between the two tropopauses. However, to our knowledge no quantitative comparison has been done regarding the role of double tropopauses in enhancing stratosphere-troposphere exchange. Quantitatively addressing this question is one of the goals of this study.

The occurrence of double tropopauses has been shown to be associated with extratropical synoptic disturbances. Studies show an eastward propagation of double tropopauses events in Hovmöller diagrams [e.g. Castanheira and Gimeno, 2011; Peevey et al., 2012], indicating their association with baroclinic Rossby waves embedded in westerly flow. Peevey et al. [2014] documented the linkage between double tropopauses and warm conveyor belts, a common feature in baroclinic disturbances. Wang and Polvani [2011] demonstrated in idealized simulations that it is the breaking of these waves that creates the largest coverage of double tropopauses. In observations this linkage between Rossby wave breaking and double tropopauses, however, has only been shown in individual case studies [e.g. Pan et al., 2009; Homeyer et al., 2014]. In this work, we present the statistical relationship between double tropopauses and synoptic Rossby wave breaking by applying a wave breaking detection algorithm to satellite observations and reanalysis data.

The specific mechanisms responsible for the formation of double tropopauses remain unknown, despite the many clues alluded to in previous studies. The clues include baroclinic wave activity and the tropopause inversion layer (TIL) [e.g. Wang and Polvani,
Figure 1. Temperature profiles from COSMIC GPS data on (a) Jan 24, 2007, and (b) Jan 22, 2007.

2011; Peevey et al., 2014], which is a thin layer with very strong stratification right above the tropopause [e.g. Birner et al., 2002; Birner, 2006]. Following these clues, we explore possible mechanisms with a focus on adiabatic synoptic processes. In particular, we propose that double tropopauses can be formed by either advection of existing tropopauses or creation of new tropopauses above old ones. We demonstrate these two mechanisms in satellite observation, reanalysis, and model simulations, and show how baroclinic waves and a TIL help form double tropopauses by advecting and creating tropopauses.

We address three questions in this study: (1) What are the statistical relationships between double tropopauses and Rossby wave breaking in the observations? (2) What are the mechanisms responsible for double tropopause formation? (3) How do double tropopauses impact stratosphere-troposphere exchange? The dataset and diagnostic methods used to address these questions are described in Section 2. Sections 3 to 5 address the three questions above in order. Section 6 summarizes the main conclusions and discusses how our results relate to those of previous studies.

2 Data and methodology

2.1 GPS radio occultation data

GPS radio occultation measurements (hereafter GPS data) provide accurate temperature observations with high vertical resolution [e.g. Liou et al., 2007], which reveal finer and more realistic tropopause structure compared to reanalysis [e.g. Birner et al., 2006;
Son et al., 2011]. We use GPS data from the Constellation Observing System for Meteorology, Ionosphere, and Climate (COSMIC) which started in late 2006. The temperature profiles were retrieved by the University Corporation for Atmospheric Research (UCAR) from 2007 to 2013. We use only dry retrievals over this period which are of the highest quality in regions with low water vapor concentration such as the tropopause layers [e.g. Wickert et al., 2005].

To study the spatial pattern of double tropopauses and their corresponding stratification, we transform thousands of temperature profiles over the entire globe each day into regularly gridded data with a horizontal resolution of $2.5^\circ \times 2.5^\circ$ and vertical resolution of 500m by linear barycentric interpolation. Such a horizontal resolution is chosen as a balance between the need for relative fine structure and a limited number of profiles. From these gridded daily temperature profiles, we calculate the potential temperature $\theta$, the Brunt-Väisälä frequency $N^2 = g\partial(ln\theta)/\partial z$, and the height of the tropopauses using the definition of the World Meteorological Organization [1957]. Specifically, the first tropopause is defined as the lowest level at which the lapse rate decreases to $2K/km$, provided that the average lapse rate between this level and any level above within $2km$ is not greater than $2K/km$. Above the first tropopause, if the average lapse rate between any level and all higher levels within $1km$ exceeds $3K/km$ again, a second tropopause is defined using the same criteria as the first.

2.2 Reanalysis data

To characterize the dynamical features of double tropopauses, we use potential vorticity (PV) from ERA-Interim reanalysis data [Dee et al., 2011] over the period from January 2007 to December 2013. The PV field has a horizontal resolution of $1.5^\circ \times 1.5^\circ$ and 15 vertical isentropic levels ranging from 265K to 850K. We apply two diagnostics to the isentropic PV field, which are described in the following subsections.

2.2.1 Equivalent latitude anomaly

Since PV contours on isentropic surfaces are material lines for adiabatic motion in the absence of friction, PV is an ideal field for characterizing advection by synoptic disturbances. Inspired by Pan et al. [2009], we first map PV ($q$) at each grid point to its corre-
Figure 2. Potential vorticity of 3 PVU (solid contour) on 350K isentropic surface and its corresponding equivalent latitude \( \phi_e \) (dashed line). The equivalent latitude anomaly \( \phi_a \) (color shading) between the 3 PVU contour and the equivalent latitude \( \phi_e \) is also plotted.

Corresponding equivalent latitude (\( \phi_e \)) value according to Butchart and Remsberg [1986]:

\[
\phi_e [q(\lambda, \phi), \theta] = \arcsin \left[ 1 - \frac{A(q, \theta)}{2\pi a^2} \right]
\]  

(1)

where \( A(q, \theta) \) is the area enclosed by a \( q \) contour on an isentrope \( \theta \) and \( a \) is the radius of the earth. By this definition, \( \phi_e (q, \theta) \) represent the latitude circle that encloses the same area as the \( q \) contour. Figure 2 shows the \( q=3 \) PVU contour on the 350K isentrope and its equivalent latitude \( \phi_e \) (dashed line). As shown in Fig. 2, an air parcel’s equivalent latitude can be thought of as the latitude it comes from. Accordingly, the difference between an air parcel’s actual latitude and its equivalent latitude:

\[
\phi_a (\lambda, \phi, \theta) = \phi - \phi_e (\lambda, \phi, \theta)
\]

(2)

can be used to measure the meridional excursion experienced by an air parcel due to synoptic adiabatic disturbances. Figure 2 illustrates the \( \phi_a \) (shading) enclosed by wave lobes associated with the \( q=3 \) PVU contour. Positive \( \phi_a \) (red shading) represents poleward advection, whereas negative \( \phi_a \) (blue shading) represents equatorward advection.

2.2.2 Rossby wave breaking detection

As shown in Fig. 2, PV contours are good at depicting the morphology of large-scale Rossby waves. In particular, wave breaking is featured in Fig. 2 by the horizontal overturning of the PV contour over the central and eastern North Atlantic. To study these breaking events and their connections to double tropopauses, we apply the PV-based wave
breaking detection algorithm described in Liu et al. [2014]. The algorithm detects wave
breaking by searching for overturning of circumpolar 2 PVU contours on isentropic sur-
faces ranging from 300K to 350K with an interval of 5K. According to the direction of
overturning, wave breaking is classified into anticyclonic wave breaking and cyclonic wave
breaking. For example, the overturning in Fig. 2 between 30°W to 20°W is cyclonic,
whereas the overturning between 20°W to 0°W is anticyclonic. For both anticyclonic and
cyclonic wave breaking events, the centroids of the overturning anticyclonic lobes are as-
signed as their centers. Throughout this paper, all composites with respect to wave break-
ing are translated so that the wave breaking centers overlap with each other. For more de-
tails, we refer readers to Liu et al. [2014].

2.3 Baroclinic eddy lifecycle simulations

One of the goals of this study is to quantify how much the occurrence of double
tropopauses enhances stratosphere-troposphere exchange compared to situations without a
double tropopause. To serve this goal, we construct a control eddy lifecycle simulation
with no double tropopause and a series of contrasting lifecycle simulations with vary-
ing areal coverage of double tropopauses following Wang and Polvani [2011] (hereafter
WP11). The advantage of these simulations is that the only difference among them is
the tropopause structure so that we can cleanly isolate any enhancement of stratosphere-
troposphere exchange due to the occurrence of double tropopauses. Note that we use a
different model from WP11, the Geophysical Fluid Dynamics Laboratory (GFDL) spec-
tral dry dynamical core with a horizontal resolution of T85, and 60 uneven sigma levels.
The initial conditions for the simulations are described in Section 2.3.1, and the setup for
passive tracers is discussed in Section 2.3.2.

2.3.1 Initial conditions

We prescribe baroclinically unstable initial conditions largely following WP11, with
some minor modifications due to the fact that we use a different model. We describe the
procedure briefly here with a focus on the differences from WP11. For the omitted details
that are common to WP11, we refer readers to the Appendices in their paper.

As in WP11, the temperature field is determined by the lapse rate field $\Gamma(\phi, z)$ which
blends a low-latitude tropical profile $\Gamma^L(z)$ with a high-latitude one $\Gamma^H(z)$. Both profiles
carry a parameter that determines the strength of the TIL: $c_L$ and $c_H$ respectively. The blending in the meridional direction takes the form:

$$\Gamma(\phi, z) = \Gamma^L(z) + \left[ \Gamma^H(z) - \Gamma^L(z) \right] \left[ \frac{1}{2} + \frac{1}{2} \tanh \left( \frac{\phi - \phi_0}{\phi_D} \right) \right]$$

(3)

which is almost the same as WP11 except that we replace metric distances $Y$, $Y_0$, and $Y_D$ with latitudes $\phi$, $\phi_0$, $\phi_D$, and set $\phi_0 = 45^\circ$, $\phi_D = 10^\circ$. The initial temperature field ($T$) is then obtained by integrating the lapse rate from the top of the atmosphere where $T = 220 K$.

With the initial temperature field defined, the zonal wind ($U$) can be obtained by utilizing the thermal wind balance on a sphere:

$$-\frac{R}{H} \frac{\partial T}{\partial \phi} = (af + 2Ut\tan \phi) \frac{\partial U}{\partial z}$$

(4)

where $R$ is the ideal gas constant, $H$ is the scale height, and $a$ is the radius of the earth. This differs from the thermal wind balance in WP11 where an $f$-plane model was used.

The zonal wind can be solved for by iterating the vertical integration of (4) as:

$$U^{i+1}(\phi, z) = \int_0^z -\frac{R}{H} \frac{\partial T}{\partial \phi} \frac{\partial U}{\partial z} \, dz$$

(5)

where we set $U(\phi, 0) = 0$. With an initial guess of $U^0(\phi, z) = 0$, the $U^i$ converges very quickly for the initial temperature field employed here (within a few iterations).

WP11 was able to increase the area of double tropopauses by increasing the strength of the initial high-latitude TIL strength $c_H$. The reason why a strong high-latitude TIL helps form double tropopause is one subject of this study and will be revisited in Section 4.2. To increase the coverage of double tropopauses, we follow WP11 except that we increase the strength of the low-latitude TIL $c_L$ by the same amount, so that the resulting jet streams have nearly the same strength under both the weak and strong TIL cases. We carry out five simulations with $c_H$ values ranging from 0K/km to 4K/km with an interval of 1K/km. For the low-latitude TIL strength, we always set it as $c_L = c_H + 3K/km$. The specific difference 3K/km is chosen to result in reasonable initial jets with a maximum zonal wind of approximately 50m/s throughout all of the five simulations. These five simulations are referred to by the high-latitude TIL strength (TIL0 through TIL4). To study the mechanisms of double tropopause formation, two of these simulations – TIL0 and TIL3 – are also used in Section 4.2.
2.3.2 Tracer setup

To quantify the influence of double tropopauses on stratosphere-troposphere exchange, we initialize two passive tracers $S$ and $T$ following Polvani and Esler [2007]. In particular, we pick the $N_{tp}^2 = 3.5 \times 10^{-4} \text{s}^{-2}$ contour as the initial tropopause. We then initialize the two tracers on isentropes ranging from 290K to 500K as:

$$S = \mathcal{H}(N^2 - N_{tp}^2)$$  \hspace{1cm} (6a)

$$T = \mathcal{H}(N_{tp}^2 - N^2)$$  \hspace{1cm} (6b)

where $\mathcal{H}(\cdot)$ is the Heaviside function:

$$\begin{cases} 
\mathcal{H}(x) = 1, & x \geq 0 \\
\mathcal{H}(x) = 0, & x < 0 
\end{cases}$$

$S + T = 1$ at all times by construction. After $t = 0$ we use the contour of $S = T = 0.5$ as the natural tropopause. The mass of $S$ in the troposphere is viewed as the accumulated stratosphere-to-troposphere ($STT$) tracer mass flux and the mass of $T$ in the stratosphere is viewed as the accumulated troposphere-to-stratosphere ($TTS$) tracer mass flux:

$$STT = \int \mathcal{H}(T - 0.5) S \rho dV$$  \hspace{1cm} (7a)

$$TTS = \int \mathcal{H}(S - 0.5) T \rho dV$$  \hspace{1cm} (7b)

3 The synoptic features of double tropopauses

Although conforming to the same WMO definition, the actual vertical temperature profiles of double tropopauses can vary substantially. Figure 1 exemplifies two types of double tropopause profiles from the COSMIC GPS data. In Fig. 1a, both the first and the second tropopauses mark a discontinuity in the thermal stratification in a similar way. Between the two tropopauses, there is a layer (13-16km in altitude) with a tropospheric lapse rate of $5-6 K/km$, similar to the lapse rate below the first tropopause. In some ways, the stratification around the second tropopause looks like a replication of that around the first tropopause. This suggests that the second tropopause in Fig. 1a may have existed before and the double tropopause was formed by horizontal advection. In Fig. 1b, the second tropopause does not mark a discontinuity, but a threshold defined by the WMO as the lapse rate decreases gradually with height back below 2K/km. In this case, the second
Figure 3. Climatology of double tropopause frequency (shading) in the Northern Hemisphere during winter for (a) total occurrence, and (b) occurrence due to differential advection (see Section 4.1 for details). Grey contours denote the climatological height of the first tropopause with an interval of 1.5km, with the largest being 15.5km.

The double tropopause seems more likely to have been created by local changes in the thermal stratification. In this section, we focus on the general features of double tropopauses first and address two possible formation mechanisms in Section 4.

Using COSMIC GPS data, the wintertime climatology of double tropopauses occurrence from 2007 to 2013 is shown in Fig. 3a (shading) along with the corresponding height of the first tropopause (contour). The double tropopauses generally occur in the subtropics within the 30°N to 40°N band where the climatological tropopause slope is most steep (grey contours). Their most frequent occurrence is located over North America and the northwest Atlantic, which is consistent with Randel et al. [2007a]. Away from the subtropics, double tropopauses also occur at higher latitudes over the North Atlantic, maximizing near the Norwegian Sea and the Greenland Sea.

To illustrate the synoptic features associated with double tropopauses, we show snapshots of two individual events in the two columns of Fig. 4. The upper row shows the meridional cross sections of thermal stratification $N^2$ (shading) and the equivalent latitude contour of $\phi_e = 30^\circ N$ derived from PV. The black dots denote the tropopauses identified by the WMO definition. The lower row shows potential temperature on the 2PVU sur-
Figure 4. The left column shows an event on Jan 24, 2017 and the right column shows an event on Feb 8, 2017. Upper row: Snapshots of the Brunt-Väisälä frequency $N^2$ (shading) from GPS temperature data and equivalent latitude contour $\phi_e=30^\circ$ (black contour) from ERA-Interim reanalysis. The black solid circles denote the tropopause identified by GPS temperature data. Lower row: Potential temperature $\theta$ (grey shading, unit: K) on the 2 PVU surface which represents the dynamical tropopause. The dotted lines denote the longitude at which those cross sections are taken and shown in the upper row. The red contours accent the overturning of $\theta$.

For the event on Jan 24 (left column of Fig. 4), the stacking of two tropopauses occurs between 25°N to 45°N, with the tropopauses vertically separated by a distance of about 6km. Corresponding to this tropopause structure, the $N^2$ field exhibits a clockwise folding. The cause of this folding is revealed by the $\phi_e = 30^\circ$ contour, which reaches 45°N at the higher tropopause level. Assuming the air
on the contour is from 30°N, this indicates a substantial poleward advection of the higher tropopause which then overlapped with the extratropical lower tropopause that remained relatively stationary. In the lower panel of Fig. 4, we show that the strong poleward intrusion at the higher tropopause level happened as a result of horizontal overturning corresponding to an anticyclonic wave breaking event.

The event on Feb 8 makes an interesting comparison with the event on Jan 24 and is shown in the right column of Fig. 4. The vertically stacked tropopauses are separated by 5-7km and a clockwise folding of N2 contour is also seen. As in the previous case, the shape of the \( \phi_e = 30°N \) contour aligns well with the N2 contour, indicating that differential advection caused the folding of the \( \phi_e = 30°N \) contour. However, the difference is that this event was dominated by an equatorward intrusion, rather than a poleward intrusion. Specifically, the air at the lower tropopause level was advected from 30°N equatorward to 20°N while the air at the higher tropopause barely moved from its equivalent latitude. The result is that the lower extratropical tropopause moved equatorward and overlapped with the tropical tropopause above. Examining the horizontal advection pattern (lower panel of Fig. 4), we also find an anticyclonic wave breaking event providing the critical equatorward intrusion. But the intrusion was associated with breaking wave’s equatorward lobe, rather than the poleward lobe.

The two examples above suggest a possible link between Rossby wave breaking and double tropopauses. We further support this linkage by compositing PV with respect to the occurrence of double tropopauses in specific regions (Fig. 5). Specifically, PV is composited for days when more than 80% of the GPS temperature profiles within the red rectangle exhibit double tropopauses. Both composites in Fig. 5 exhibit large-scale horizontal overturning of PV contours indicative of Rossby wave breaking. In particular, the double tropopauses over the subtropical northeastern Pacific tend to occur in the equatorward lobe of anticyclonic wave breaking (Fig. 5a), whereas those over the Labrador Sea tend to occur in the equatorward lobe of a cyclonic wave breaking (Fig. 5b).

As Rossby wave breaking is shown to be linked to double tropopauses climatologically, we ask “what are the mechanisms through which they are linked”. The snapshots in Fig. 4 suggest that wave breaking may help form double tropopauses by providing vertically differential advection, either stronger poleward advection of the higher tropical
Figure 5. Composite potential vorticity (PV) in unit of PVU on isentropes for days when more than 80% of profiles within the red rectangle exhibit double tropopauses.
tropopause or stronger equatorward advection of the lower extratropical tropopause. We will further investigate this hypothesis in the next section.

4 Mechanisms for the adiabatic formation of double tropopauses

Since double tropopauses are defined by thermal stratification, looking at its budget should provide clues to the formation of double tropopauses. Below is the adiabatic prognostic equation of $N^2$ using the approximation of thermal wind balance for large-scale flow:

$$\frac{\partial N^2}{\partial t} = -\mathbf{V} \cdot \nabla N^2 + N^2 \frac{\partial w}{\partial z}$$

The terms on the RHS correspond to two physical processes that contribute to the local change in $N^2$, as alluded to in Fig. 1. The first is the advection term (A) and the other is the compression/stretching term (B) that produces/destroys $N^2$. When the production term B is small, $N^2$ behaves like a passive tracer and the formation of double tropopause can be mainly attributed to differential horizontal advection. When the production term is not small, double tropopauses can be formed even in the absence of differential advection. We explore these two mechanisms in the next two subsections.

While the evolution of $N^2$ is likely always a mixture of both A and B, we aim to understand the behavior of tropopause evolution under each separately as a stepping stone to understanding more complex scenarios. In addition, we illustrate in observations and model simulations that the formation of double tropopause can at times be predominantly determined by one of these two mechanisms.

4.1 Differential advection

Snapshots in Fig. 4 illustrate a possible mechanism by which Rossby wave breaking helps form double tropopauses – providing differential advection. To examine how robust this mechanism is in climatology, we composite variables for the two types of Rossby wave breaking events in Fig. 6. The upper row shows the composite double tropopause frequency anomaly relative to the local climatology (shading) and the lower row shows the corresponding thermal stratification and advection pattern. The composites are made by horizontally shifting the fields about the wave breaking centers, so the x and y axes are relative longitude and latitude respectively. We only use centers with a latitude between
30°N and 55°N, to focus on the region where the climatological tropopause is steep and differential advection is most likely to operate.

Figure 6. Composite variables for anticyclonic wave breaking (AWB) in the left column and cyclonic wave breaking (CWB) in the right column. The fields are moved horizontally so that the centers for each type of wave breaking overlap. The longitude and latitude shown on axes are relative to the wave breaking centers.

Upper row: Anomalous frequency of double tropopause occurrence (shading) and potential temperature θ (contour) on the 2 PVU surface which represents the dynamical tropopause. Lower row: Cross section of N² (contour) taken along the dashed lines in the upper row. The color shading is added to the folding contours and denotes the equivalent latitude anomaly φₐ. The interval of the N² contours is 0.5 × 10⁻⁴ s⁻², and the value for the two colored contours are 3.5 × 10⁻⁴ s⁻², 4 × 10⁻⁴ s⁻² respectively.

Double tropopauses occur significantly more frequently than climatology to the north of anticyclonic wave breaking (Fig. 6a). Higher frequencies of double tropopauses
also occur to the south of anticyclonic wave breaking’s equatorward lobe, but with much weaker strength. To visualize the vertical structure of stratification associated with the anomalous double tropopause formation shown in Fig. 6a, we plot a meridional cross section of $N^2$ (contours) along $-17.5^\circ$ relative longitude in Fig. 6c. The $N^2$ contours fold substantially in the composite, demonstrating the robustness of this double tropopause structure. To see if the folding is related to differential advection, we color the folding contours by their anomalous equivalent latitude $\phi_a$. The red shading indicates poleward advection and the blue shading indicates equatorward advection. The folding in Fig. 6c is mainly associated with extensive poleward advection at 16km height and little advection near 13.5km, depicting a major role for differential advection in the formation of these double tropopauses in a composite sense. Comparing Fig. 6a and Fig. 6c, one can see the poleward advection north of -10° relative latitude is provided by the poleward advection of anticyclonic wave breaking, as is also exemplified in the left column of Fig. 4.

For cyclonic wave breaking, the largest positive anomaly of double tropopause frequency occurs at the south edge of its equatorward lobe (Fig. 6b). To visualize the meridional structure of thermal stratification associated with the largest anomaly, another cross section is taken along $-2.5^\circ$ relative longitude and shown in Fig. 6d. Similar to anticyclonic wave breaking, $N^2$ contours fold substantially, indicating the robustness of the double tropopause anomalies. What is different, however, is that the folding is dominated by equatorward advection of the lower tropopause near the height of 11km. The dominating equatorward advection to the south of -5° relative latitude is provided by the equatorward lobe of cyclonic wave breaking. Comparing Fig. 6c with Fig. 6d, it is evident that the double tropopauses formed in the cyclonic lobe by equatorward advection are more vertically separated than those formed in the anticyclonic lobe by poleward advection, which is also alluded in Figs. 4a and 4b. This is consistent with Wirth [2001] who studied the double tropopauses resulting from idealized differential advection of PV.

The composites for wave breaking along with snapshots in Fig. 4 demonstrate that differential advection can dominate the formation of double tropopauses in a composite sense. We proceed by addressing how often this happens, regardless of the occurrence of wave breaking. To answer this, we classify each double tropopause occurrence into advective and non-advective using the equivalent latitude $\phi_e$ field. In words, for each second tropopause, we test if it still overlaps with a lower tropopause after we move all the
tropopauses meridionally to their equivalent latitudes. If it does, we classify it as a non-adveective double tropopause. Otherwise we mark it as advective.

The result of the classification is shown in Fig. 3b. Advective double tropopause occurrence is more confined in the tropopause break region than the total occurrence. This supports the hypothesis that the advective mechanism is most likely to operate in the tropopause break region. In particular, the most frequent occurrence of advective double tropopauses is located over east Asia and the North Pacific near Japan where the tropopause slope is steepest. Summing up the total occurrence of advective double tropopauses, differential advection appears to account for 47% of the total occurrence of double tropopauses in the Northern Hemisphere.

Besides differential advection, previous studies have also focused on the direction of the mean advection above the first tropopause [e.g. Pan et al., 2009; Añel et al., 2012; Schwartz et al., 2015; Wang and Polvani, 2011]. We next quantify how favorable various advection patterns are for double tropopause formation. Specifically, we take all the GPS profiles and calculate the frequency of double tropopause as a function of both mean advection ($\phi_{mean}^a$) and differential advection ($\phi_{diff}^a$) associated with the profiles (Fig. 7). Positive and negative $\phi_{mean}^a$ indicates poleward and equatorward mean advection respectively. Positive and negative $\phi_{diff}^a$ indicate clockwise and counterclockwise folding respectively. The four corners of Fig. 7 correspond to the four combined advection patterns which are illustrated by schematics at their corresponding corners. For detailed definitions of $\phi_{mean}^a$ and $\phi_{diff}^a$, we refer readers to Appendix A.

In Fig. 7, the occurrence of double tropopauses is largely dictated by differential advection $\phi_{diff}^a$. That is, the frequency of double tropopauses is mainly stratified in the vertical direction of the plot, becoming increasingly more frequent as the value of $\phi_{diff}^a$ increases for any $\phi_{mean}^a$ bin. Particularly large increases occur near the transition from negative $\phi_{diff}^a$ to positive $\phi_{diff}^a$, the latter of which represents the favorable shear direction for forming double tropopauses. When the shear direction is in favor of double tropopauses ($\phi_{diff}^a > 0$), poleward mean advection ($\phi_{mean}^a > 0$) is more likely to form double tropopauses than equatorward mean advection ($\phi_{mean}^a < 0$). Physically, this means that double tropopauses are more likely to be formed by the poleward advection of the higher tropopause than the equatorward advection of the lower tropopause. This is consistent with Castanheira and Gimeno [2011] where they found the poleward edge of double
Figure 7. The frequency of wintertime double tropopauses for COSMIC GPS profiles binned by mean equivalent latitude anomaly $\phi^\text{mean}_a$ (x-axis) and the vertical shear of equivalent latitude anomaly $\phi^\text{diff}_a$ (y-axis). $\phi^\text{mean}_a$ and $\phi^\text{diff}_a$ are calculated from ERA-Interim reanalysis. The meridional advection patterns corresponding to the four corners are illustrated by schematics, with red lines denoting the tropopauses and black arrow denoting the meridional wind.
Figure 8. Snapshots of the Brunt-Väisälä frequency $N^2$ (shading) from GPS temperature data and equivalent latitude contour $\phi_e=35^\circ$ (black contour) from ERA-Interim reanalysis on Jan 22, 2007. The black solid circles denote the tropopause identified by GPS temperature data.

tropopause area is better correlated with the meridional extent of double tropopauses than the equatorward edge. The preferred poleward advection of tropospheric air is also suggested by Castanheira et al. [2012] where lower total column ozone is found for double tropopauses. Since ozone is rich in the stratosphere, lower column ozone indicates more tropospheric air in the column and hence suggests a poleward tropospheric intrusion into the stratosphere.

4.2 Destabilization/vertical stretching

In the previous section we showed that approximately half of double tropopauses are formed directly by differential advection. In this section we explore possible mechanisms for the formation of the other 50%. Figure 8 shows a cross section of a double tropopause event that is not directly caused by differential advection. The stacking of tropopauses occurs between 27°N and 38°N where the $N^2$ field (shading) shows a clockwise folding.
However, the equivalent latitude ($\phi_e$) contour of $30^\circ$N (contour) shows a counterclockwise folding. This contrasts with the cases shown in Figs. 4 where the $\phi_e$ contour approximately aligned with the $N^2$ contours, indicating that $N^2$ behaved largely as a passive tracer. In Fig. 8, the non-passive behavior of $N^2$ indicates that production/destruction processes of thermal stratification might be responsible for this double tropopauses event, such as the vertical stretching term $B$ of (8).

Since stretching/compression is the only destruction/production process for adiabatic $N^2$, we quantify the accumulated effect of this stretching/compressing process by comparing $N^2$ with its passive counterpart in adiabatic situations. In particular, we conduct adiabatic eddy lifecycle simulations as described in Section 2.3 and initialize a passive tracer $N^2_P$ that has the same initial value as $N^2$. Then the difference $\Delta N^2 = N^2 - N^2_P$ quantifies the accumulated destruction/production of $N^2$, namely the effect of stretching/compression that destabilizes/stabilizes the thermal stratification. We further normalize $\Delta N^2$ by $N^2$ to get the fractional change of $N^2$ due to stretching/compression.

The quantification of destabilization/stabilization and its role in forming double tropopauses during the eddy lifecycle simulations is summarized in Fig. 9. The left column shows a simulation with no TIL (hereafter TIL0) and the right column shows a simulation with a TIL of strength $c_H = 3K/km$ (hereafter TIL3). The upper row shows that TIL3 has double tropopauses in the cyclonic lobe of wave breaking (red shading) whereas TIL0 does not. Note that the identification of double tropopauses is the same as in WP11, replacing the 3K/km criteria in the WMO by 2K/km. Also shown in the upper row is $N^2_P$ (contour) which depicts the horizontal advection pattern. Despite the difference in double tropopause coverage between TIL0 and TIL3, there is no obvious reason that this difference is due to difference in the horizontal advection pattern. This suggests that TIL may influence the formation of double tropopauses through some other mechanisms than horizontal advection.

The middle and bottom rows of Fig. 9 show cross sections of $N^2_P$ (dashed contour) on day 1 and day 8. The tropopause ($N^2 = 3.3 \times 10^{-4}s^{-2}$) is highlighted for both $N^2_P$ (dashed magenta) and $N^2$ (solid magenta), so that the solid magenta contour denotes the actual tropopause and the dashed magenta contour denotes the hypothetical tropopause if there were no stretching/compression. The two tropopause contours are same on day 1 but differ on day 8 in the tropopause break region between 40°N and 50°N. The frac-
Figure 9. **Left column:** (a) Concentration of $N_P^2$ tracer (contour) and the location of double tropopauses (red shading) in the simulation without an extratropical TIL. (c) Cross section of $N_P^2$ (dashed contour) and $N^2$ (solid line) along the dashed line shown in (a) on day 1. The magenta contours have a value of $3.3 \times 10^{-4} \text{s}^{-2}$ and represents the thermal tropopause. Black contours have an interval of $0.5 \times 10^{-4} \text{s}^{-2}$. (e) The contours are the same as in (c) but for day 8. The shading is the fractional material change in thermal stratification $\Delta N^2/N_P^2$. **Right column:** The same as the left column, except for the simulation with an extratropical TIL.
Figure 10. Accumulated tracer mass flux between the stratosphere (S) and the troposphere (T) in idealized eddy lifecycle simulations as a function of (a) area fraction of double tropopauses coverage, and (b) area fraction of the tropopause. The area fraction is a unitless value defined as area divided by the global surface area.

...tional stretching/compression ($\Delta N^2/N_P^2$) that leads to such a change is plotted as shading in the bottom row. In particular, for both TIL0 and TIL3 it is the stretching (blue shading in Figs. 9e-f) in the tropopause break region that destabilizes the thermal stratification and hence causes the actual thermal tropopause to deviate from its hypothetical counterpart. However, the specific deviations are not the same for TIL0 and TIL3, as the former does not increase tropopause folding whereas the latter does. Interestingly, this is not due to the stretching ($\Delta N^2$) but the initial $N^2$ that $\Delta N^2$ adds to. In the case of TIL0, the stretching/destabilization destroyed the initial tropopause and created a new one above it between 40°N and 50°N, so that the net effect was to move the tropopause upward (Fig. 9e). In the case of TIL3, while the stretching created a new tropopause at higher levels, it was not able to destroy the lower initial tropopause because of the strong initial stratification ($N^2$) (Figs. 9d and 9f). The result is the stacking of two tropopauses between 40°N and 45°N. Due to these differences in the initial stratification – with or without a TIL – destabilizations of similar strength and spatial pattern lead to very different modification of the thermal tropopause.
5 Enhancement of stratosphere-troposphere exchange

Both double tropopauses and wave breaking are important processes that enhance the stratosphere-troposphere exchange by disturbing the tropopause interface. While detailed quantification of exchange during idealized wave breaking simulations was performed by Polvani and Esler [2007], the impact of double tropopauses on stratosphere-troposphere exchange has not been quantified to the best of our knowledge. Here, we ask to what extent the occurrence of double tropopauses enhances the stratosphere-troposphere exchange during wave breaking events. To answer this question, we use the same setting as in the previous section to generate idealized model simulations with various extents of double tropopauses. In particular, we simulate five eddy lifecycles (TIL0-TIL4) featuring initial TIL strength ranging from 0 K/km to 4 K/km (see Section 2.3.1 for details), with TIL0 and TIL3 also being used to study the formation mechanisms of double tropopause in Section 4.2. Then we apply the method of Polvani and Esler [2007] to quantify the corresponding strength of the stratosphere-troposphere exchange. In words, we quantify the amount of stratospheric tracer in the troposphere and the amount of tropospheric tracer in the stratosphere as the two-way mass exchange across the tropopause (see 2.3.2 for details). The results are plotted in Fig. 10a as a function of double tropopause area fraction (the coverage area divided by global surface area $4\pi a^2$). As the area fraction of double tropopauses increases, the troposphere-to-stratosphere exchange stays largely unchanged but the stratosphere-to-troposphere exchange increases substantially, as much as double the magnitude when no double tropopause is present.

One way double tropopauses may amplify the stratosphere-to-troposphere exchange during wave breaking is by increasing the area of the tropopause by vertical folding. As the tropopause area increases, there exists a larger interface between the stratosphere and the troposphere for transport and mixing to occur and thus, this might lead to more exchange of air mass. We test this by calculating the total tropopause area ($A_{TP}$) as the area of the isosurface $S=T=0.5$ during these experiments. We then plot the mass exchange as a function of $A_{TP}/4\pi a^2$ in Fig. 10b. As the tropopause area increases, the troposphere-to-stratosphere exchange stays largely unchanged but the stratosphere-to-troposphere exchange increases significantly, similar to Fig. 10a. This provides evidence that the increase in the tropopause area is the mechanism by which double tropopauses enhance the stratosphere-troposphere exchange. Combining this with the results shown in Fig. 9b, we argue that the
increased tropopause area by double tropopauses in wave breaking lobes can substantially enhance the strength of stratosphere-to-troposphere exchange.

6 Conclusions and discussion

Using COSMIC GPS data, ERA-Interim reanalysis, and eddy lifecycle simulations of the GFDL dynamical core, we investigate the synoptic features and transport/mixing strength of double tropopauses, as well as the adiabatic mechanisms responsible for their formation. In particular, we address three questions with their summarized answers below:

1. What is the relationship between double tropopauses and Rossby wave breaking in the observations?

   In a composite sense, anticyclonic and cyclonic wave breaking are both linked to increased double tropopause occurrence (Fig. 6), and double tropopause occurrence in certain regions also shows the signature of both types of wave breaking (Fig. 5).

2. What are the mechanisms responsible for double tropopause formation?

   We propose two mechanisms. The first one is differential advection of thermal stratification that leads to the folding of the thermal tropopause (Fig. 4). In particular, Rossby wave breaking helps form double tropopauses by providing such differential advection in the observations (Fig. 6). Climatologically, approximately 50% of all the double tropopauses are formed by this mechanism (Fig. 3). The second mechanism is the creation of the tropopause by destabilizing the thermal stratification in the presence of a TIL (Fig. 9).

3. How do double tropopauses impact stratosphere-troposphere exchange?

   In idealized eddy lifecycle simulations, the occurrence of double tropopauses substantially enhances the stratosphere-to-troposphere exchange (Fig. 10), whereas the troposphere-to-stratosphere exchange is largely unchanged. We suggest that the enhancement is due to increased tropopause area corresponding to the folding tropopause.

   Previous research suggested that cyclonic flow is preferred near double tropopauses [e.g. Randel et al., 2007a; Wang and Polvani, 2011]. The destabilization/stretching mechanism can potentially explain this preference. When an air parcel is adiabatically stretched/destabilized, the pseudo-density \( \sigma = \frac{1}{\gamma} \frac{\partial \rho}{\partial \theta} \) increases. To conserve \( PV = (f + \zeta) / \sigma \), absolute vorticity \( f + \zeta \) has to increase. For baroclinic waves, this increase is a result of an increase in
relative vorticity $\zeta$ rather than $f$ (not shown). Therefore, relative vorticity is likely to be positive in the presence of adiabatic stabilization.

In the eddy lifecycle simulations, double tropopauses only enhance the equatorward stratosphere-to-troposphere exchange, which is inconsistent with the dominant poleward advection associated with double tropopauses in the observations. The reason for the equatorward enhancement in our idealized simulation may be that the double tropopauses only occur in the equatorward lobe of the breaking wave where the exchange across the tropopause is dominated by an equatorward flux (not shown). However, in observations double tropopauses occur in both poleward and equatorward lobes of wave breaking (Fig. 6) and poleward advection is more favorable for double tropopauses [e.g. Randel et al., 2007a; Pan et al., 2009; Castanheira and Gimeno, 2011]. Therefore, double tropopauses in the observations may actually predominantly enhance poleward mixing of tropospheric air into the stratosphere. More research is needed to test this.

The dominant mechanism responsible for double tropopause formation during Rossby wave breaking events also appears to be different between the eddy lifecycle simulations and the observations. In particular, differential advection contributes to the dominant double tropopause formation near the poleward lobe of anticyclonic wave breaking while in the eddy lifecycle simulation vertical stretching contributes to the double tropopause formation in its equatorward lobe. We suggest such difference in the wave breaking characteristics may be related to very different jet configurations in the real atmosphere and in the idealized lifecycle simulation.

Our study focuses on adiabatic mechanisms of double tropopause formation by synoptic waves. However, there are certainly other mechanisms that are relevant. One example is mean advection. Specifically, we depict the advection we are interested in by equivalent latitude anomaly $\phi_a$, which excludes the influence of mean advection. Birner [2010] showed that the climatological mean meridional structure of $N^2$ exhibits slight folding in the subtropics and mean advection contributes to this folding. Another component we neglect is the diabatic change in thermal stratification by radiative processes involving ozone and water vapor. Studies have shown their role in forming a TIL [e.g. Randel et al., 2007b; Ferreira et al., 2016; Kunkel et al., 2016]. It is possible these diabatic processes can influence the formation of double tropopauses directly, or indirectly by their ability to modify TIL. In addition, as ozone and water vapor distributions near the tropopause are
mainly determined by transport and mixing, their feedbacks on thermal stratification are likely coupled to adiabatic processes such as differential advection. More research is warranted to unfold these interactions.

Among the results, perhaps the most striking one is how much double tropopauses can enhance the stratosphere-to-troposphere exchange in idealized eddy lifecycles which are themselves strong in transport and mixing to begin with [e.g. Polvani and Esler, 2007]. Given the ubiquitousness of double tropopauses in the observations, they likely play a significant role in the stratosphere-troposphere exchange and account for its variability at least on synoptic time scales. Our results regarding their linkage to wave breaking and their formation mechanisms may then help advance understanding of the variability of the global stratosphere-troposphere exchange.

A: Quantification of $\phi_{a}^{diff}$ and $\phi_{a}^{mean}$

To quantify the advection patterns corresponding to each GPS profile, ERA-Interim data is used to calculate the equivalent latitude anomaly profile $\phi_{a}(\theta)$ collocated with the GPS profiles. By summing and subtracting $\phi_{a}$ in the vertical direction, we isolate two aspects of the advection patterns – differential advection ($\phi_{a}^{diff}$) and mean advection ($\phi_{a}^{mean}$), respectively. Differential advection represents the shear of advection in meridional plane and is calculated as:

$$\phi_{a}^{diff} = \frac{\phi_{a}(380K) - \phi_{a}(\theta_{TP})}{380K - \theta_{TP}} \quad (A.1)$$

where $\theta_{TP}$ is the potential temperature $\theta$ at the first tropopause. 380K is used here for an upper level above the first tropopause. The result is not sensitive as we vary this value between 380K and 450K (not shown).

The mean advection represents the average advection direction above the lowest tropopause and is calculated as:

$$\phi_{a}^{mean} = \frac{\phi_{a}(380K) + \phi_{a}(\theta_{TP})}{2} \quad (A.2)$$

Positive values denote poleward mean advection and negative values denote equatorward mean advection.
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References


Birner, T., A. Dörnbrack, and U. Schumann (2002), How sharp is the tropopause at midlatitudes?, *Geophysical research letters*, 29(14).


