Transport and Mixing in the Extratropical Tropopause Region in a High-Vertical-Resolution GCM. Part II: Relative Importance of Large-Scale and Small-Scale Dynamics

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ABSTRACT

The relative roles of atmospheric motions on various scales, from mesoscale to planetary scale, in transport and mixing in the extratropical tropopause region are investigated using a high-vertical-resolution general circulation model (GCM). The GCM with a vertical resolution of about 300 m explicitly represents the propagation and breaking of gravity waves and the induced transport and mixing. A downward control calculation shows that the Eliassen–Palm (E-P) flux of the gravity waves diverges and induces a mean equatorward flow in the extratropical tropopause region, which differs from the mean poleward flow induced by the convergence of large-scale E-P fluxes. The diffusion coefficients estimated from the eddy potential vorticity flux in tropopause-based coordinates reveal that isentropic motions diffuse air between 20 K below and 10 K above the tropopause from late autumn to early spring, while vertical mixing is strongly suppressed at around 10–15 K above the tropopause throughout the year. The isentropic mixing is mainly caused by planetary- and synoptic-scale motions, while small-scale motions with a horizontal scale of less than a few thousand kilometers largely affect the three-dimensional mixing just above the tropopause. Analysis of the gravity wave energy and atmospheric instability implies that the small-scale dynamics associated with the dissipation and saturation of gravity waves is a significant cause of the three-dimensional mixing just above the tropopause. A rapid increase in the static stability in the tropopause inversion layer is considered to play an important role in controlling the gravity wave activity around the tropopause.

1. Introduction

In a companion paper (Miyazaki et al. 2010, hereafter Part I) we examined the roles of dynamic and thermodynamic processes in maintaining the extratropical transition layer (ExTL) and the tropopause inversion layer (TIL). A large vertical temperature gradient in the lowermost stratosphere forms the TIL (Birner et al. 2002), while chemical constituent concentrations in the ExTL are intermediate to those typical in the troposphere and stratosphere (Fischer et al. 2000; Hoor et al. 2004). These two layers are known to be almost collocated just above the extratropical tropopause (e.g., Heglin et al. 2009). The analysis used a vertically highly resolved general circulation model (GCM) that had a horizontal resolution of T213 and a vertical resolution of about 300 m above the extratropical upper troposphere. The high-resolution GCM was capable of realistically simulating atmospheric finescale structures in the extratropical upper troposphere and lower stratosphere (UTLS), including the ExTL and TIL. From a potential vorticity (PV) and thermodynamic balance analysis, we found that the ExTL and TIL can have similar locations as a result of common dynamic processes and interactions between their constituent distributions and thermal structures in the extratropical tropopause region—that is, downward advection of constituent concentration gradients and heat at the lower part of these layers, and the mixing of constituents (including water vapor) and radiative stratification effects related to large constituent concentration
gradients at the upper part of these layers. Although Part I provided insight into the atmospheric processes in the extratropical tropopause region, their relative importance at different spatiotemporal scales has not yet been discussed.

Planetary-scale and synoptic-scale waves are known to be important in determining atmospheric structures and constituent distributions in the extratropical UTLS. Planetary-scale waves cause significant mixing within the stratospheric surf zone (McIntyre and Palmer 1984) and induce a poleward and downward mean-meridional circulation (i.e., Brewer–Dobson circulation) in the extratropical stratosphere (e.g., Rosenlof and Holton 1993). The mean downward motion induced by planetary waves provides an important transport pathway to the extratropical tropopause region from higher levels. Synoptic-scale waves (e.g., baroclinic waves) cause significant transport and mixing in the upper troposphere and around the tropopause in the extratropics (e.g., Stone et al. 1999). The exchange of air caused by these large-scale motions across the tropopause plays an essential role in the abundance of chemical constituents in both the troposphere and stratosphere (Holton et al. 1995): for example, the stratosphere–troposphere exchange (STE) on the isentropic surface is mainly due to mixing by Rossby wave breaking and stirring effects by differential advection near the jet stream (e.g., Chen et al. 1994; Postel and Hitchman 1999, 2001). Moreover, medium-scale waves, with a typical zonal wavelength of 2000–3000 km, also affect the circulation near the midlatitude tropopause. Active medium-scale waves have been observed in spring at slightly higher latitudes than the subtropical jet stream (Sato et al. 1993; Yamamori et al. 1997).

In addition to large-scale and medium-scale motions, small-scale motions by turbulence and gravity wave breaking play important roles in transporting air around the extratropical tropopause (e.g., Clark and Peltier 1977; Shapiro 1980; Lane et al. 2003; Pavelin et al. 2002). Gravity wave saturation processes play roles in determining the spectrum of atmospheric motions and lead to small-scale motions. Gravity wave breaking results in three-dimensional (3D) turbulence and mixing (e.g., Fritts and Alexander 2003). Gravity wave momentum fluxes act to accelerate (or decelerate) large-scale flows at levels where gravity waves are transient or dissipating through wave–mean flow interactions. Turbulence dominates very small-scale motions around the tropopause, particularly at scales of less than several kilometers (Duck and Whiteway 2005). Observational studies have found strong gravity wave activity around and just above the tropopause (e.g., Allen and Vincent 1995; Lamarque et al. 1996; Lane and Sharman 2008; Schmidt et al. 2008). Lamarque et al. (1996) demonstrated that the topographically excited gravity waves that break at the tropopause level are responsible for irreversible mixing across the tropopause over a mountainous region. Lane and Sharman (2008) showed that gravity wave breaking dominates turbulent generation above deep convection in which the critical level of this gravity wave breaking is controlled by the above-cloud wind shear. At the same time, an important source of inertia–gravity waves is located near the tropopause level at atmospheric jets (e.g., Plougonven and Snyder 2005, 2007). Rapid changes in the static stability around the TIL can obviously affect the small-scale dynamics related to the propagation and breaking of gravity waves (e.g., vanZandt and Fritts 1989). The small-scale dynamics may provide significant contributions to the transport and mixing processes in the extratropical tropopause region.

High-vertical-resolution GCMs with no gravity wave drag parameterizations can explicitly represent the propagation and breaking of gravity waves and the induced wave–mean flow interactions. In this paper, we analyze the output of a high-vertical-resolution GCM to demonstrate the relative importance of atmospheric motions with various scales, from the gravity wave scale to planetary scale, on transport and mixing in the extratropical tropopause region. The structure of this paper is as follows: In section 2, we describe a high-vertical-resolution GCM and the analysis methods for the downward control calculation and diffusion coefficients. Section 3 presents the relative roles of waves of various scales in driving the mean meridional circulation. In section 4, we present the mixing characteristics in the extratropical tropopause region and the relative contributions of atmospheric motions with different spatial scales to this mixing. In section 5, we discuss gravity wave activity in the extratropical tropopause region. Conclusions are given in section 6.

2. Methodology

a. Description of a T213L256 GCM

We adopted a high-vertical-resolution GCM developed by Watanabe et al. (2008) to investigate the transport and mixing processes in the extratropical UTLS. This GCM has been used to survey various small-scale phenomena, including gravity waves and their role in large-scale circulation fields in the atmosphere (Watanabe et al. 2008, 2009; Tomikawa et al. 2008; Part I; Sato et al. 2009). The GCM has a T213 truncation in the horizontal (i.e., latitude–longitude grid intervals of 0.5625°) direction and 256 levels in the vertical direction, from the surface to about 85 km, with a vertical resolution of about 300 m above the extratropical upper troposphere through the
middle atmosphere. The aspect ratio of the horizontal to vertical resolution (~300) may be very appropriate for the UTLS as discussed in Birner (2006). Gravity waves are explicitly simulated in this high-vertical-resolution GCM, with no gravity wave drag parameterizations, for both orographic and nonorographic components. The high vertical resolution allows us to resolve a large portion of the observed gravity waves; their typical vertical wavelengths are 1–4 km in the lower stratosphere (Sato 1994; Sato et al. 1997). Although the high-vertical-resolution GCM makes it possible to study various small-scale phenomena including gravity waves, it still misses a great deal of the GW spectrum (scales smaller than a few hundred kilometers).

Part I demonstrated that the high-vertical-resolution GCM simulated the thermal and dynamical finescale structure around the extratropical tropopause region reasonably well. The thickness and maximum stability of the TIL was much better represented in the high-vertical-resolution GCM than in a standard-resolution GCM (with a horizontal resolution of T42 and 32 vertical levels from the surface to the upper stratosphere). Further details of the GCM are given in Watanabe et al. (2008). The results of the analysis using hourly GCM output are presented as fields averaged over three years.

b. Analysis methods

1) DOWNWARD CONTROL CALCULATION

The Eliassen–Palm (E-P) flux divergence drives the mean meridional circulation through wave–mean flow interactions. The relationship can be diagnosed from the downward control principle (Haynes et al. 1991) for steady-state conditions in the transformed Eulerian mean (TEM) framework (Andrews and McIntyre 1976):

$$ \psi = \int^z \left( \frac{\rho_0 a^2 F \cos^2 \phi}{\bar{m}} \right)_{\phi=\phi(z')} \, dz', $$

$$ \overline{\bar{w}^*} = -\frac{1}{\rho_0 \cos \phi} \frac{\partial \psi}{\partial z}, $$

where $\psi$ is the streamfunction, $\overline{\bar{w}^*}$ is the mean vertical velocity, $F$ is the divergence of the E-P flux estimated in the TEM framework, $a$ is the earth’s radius, $\rho_0 = \exp(-z/H)$ is the atmospheric density, $H$ is the scale height (~7 km), $\phi$ is the latitude, and $z$ is the pressure height. The vertical integration is along a line of constant angular momentum, $\bar{m} = a \cos \phi (\pi + a \Omega \cos \phi)$; $dz'$ is the vertical projection of a segment of $\bar{m}$. The relationship in the downward control principle expresses the fact that the interaction between the waves and mean flow is the driving force behind the mean meridional circulation, and this is in turn related to the nonacceleration theorem.

The adjustment time for the wave–mean flow interactions is several weeks in the lower stratosphere (Haynes et al. 1991); thus, it is possible to apply the downward control calculation to estimate the mean meridional circulation by considering sufficiently long time averages (Rosenlof and Holton 1993). The downward control calculation was performed in $z$ coordinates, and then the analysis results were converted into isentropic coordinates using zonal mean geopotential height and potential temperature fields.

By using a zonal mean mass continuity equation, the mean meridional wind velocity is estimated as

$$ \overline{u^*} = -\frac{1}{\rho_0 \cos \phi} \frac{\partial \psi}{\partial z}. $$

To estimate the relative contributions of the wave phenomena on various scales driving the mean-meridional circulation, the E-P flux is separated into three groups: planetary waves (PWs), medium-scale waves (MWs), and gravity waves (GWs). The PW group is defined as zonal wavenumber ($s$) components 1–3 using fast Fourier transformation (FFT). The MW group is defined as the total horizontal wavenumber ($n$) components 1–20, excluding the $s = 1–3$ components, which are extracted with a spherical filter using the Legendre transformation. The MW group includes synoptic-scale waves and subsynoptic (medium)-scale waves. The GW group is mostly composed of gravity waves, which are extracted with a spherical high-pass filter. The horizontal wavelengths of the GW group are between 188 and 1900 km; these are not represented by most climate models (i.e., low-resolution GCMs). Watanabe et al. (2008) confirmed that the high-vertical-resolution GCM can resolve atmospheric waves with horizontal wavelengths of at least 210 km. It should be noted that the GW group includes both gravity waves and subsynoptic-scale eddies.

2) DIFFUSION COEFFICIENTS

We can characterize atmospheric mixing processes by estimating the diffusion coefficient using the eddy PV flux. By assuming a flux–gradient linear relationship, the diffusion coefficient provides a good measure of atmospheric mixing. The meridional diffusion coefficient in isentropic coordinates $K_{yy}$ can be derived from the eddy meridional flux and meridional gradient of PV on isentropic surfaces (Tung 1986; Bartels et al. 1998; Newman et al. 1988; Miyazaki and Iwasaki 2005):

$$ \left[ \overline{u'q'} \right]_t \simeq -K_{yy}(t) \left( \frac{\partial \bar{q}^*}{\partial \phi} \right)_t, $$
where \( q \) is the PV, \( \nu \) is the meridional wind, \( \theta \) is the potential temperature, and \([-\cdot\cdot]_t\) denotes the time average. The asterisks and overbars represent mass-weighting and isentropic zonal means, respectively. We assume that the source and sink terms of the PV due to diabatic processes and slant diffusions \( \{-K_y(t)[(\partial q/\partial z)_{y,0}\}\} \) are negligible. Lyjak and Yudin (2005) demonstrated that the diagnostic results for eddy mixing estimated from \( K_{yy} \) agree well with those estimated from the equivalent mixing length (whose square value is proportional to the effective diffusivity; Nakamura 1996) in the extratropical UTLS. Part I also noted that most of the mixing properties are the same for \( K_{yy} \), as estimated from the PV and from long-lived tracers in the extratropical UTLS. This commonality suggests that the PV can be used as a tracer to diagnose atmospheric mixing. In addition, Part I confirmed that the diabatic source–sink effect on the PV budget is much less significant than transport effects in the extratropical UTLS. Therefore, the analysis results calculated with the eddy PV flux can be used to investigate atmospheric transport characteristics in these regions.

Steady conservative wave motions projected onto a meridional plane cause apparent diffusion in addition to the true diffusion caused by dissipative wave motions in \( K_{yy} \) (Miyazaki and Iwasaki 2005). To remove the apparent diffusion effect caused by conservative wave motions and represent the \( K_{yy} \) caused by the true mixing, similar to the effective diffusivity, a time average window \( t \) can be applied to the eddy PV flux. Time-averaged eddy PV flux fields act to reduce the effect of the apparent diffusion on \( K_{yy} \). Here \( t \) is set to 1 day based on several sensitivity calculations. By increasing \( t \) to longer than 1 day, \( K_{yy} \) became only slightly smaller, but its spatial structure was somewhat noisier because of the smaller number of averaged samples (not shown).

A vertical diffusion coefficient derived from a flux–gradient relationship with the eddy vertical PV flux may not provide a good measure of atmospheric mixing. This is because variations in static stability lead to nonlinearity in the flux–gradient relationship. Instead, we can measure the strength of the vertical mixing from the vertical (diabatic) component of the eddy PV flux normalized by a background PV value:

\[
\frac{\mathcal{Q}'q'}{q}
\]

where \( \mathcal{Q} \) is the diabatic heating rate that leads to vertical air movements in isentropic coordinates. Thus, the vertical (diabatic) component of the eddy flux represents the strength of the vertical dispersions that deviate from isentropic surfaces through eddy motions. The eddy vertical PV flux increases as the PV increases with height. To measure the vertical mixing strength independent of the PV value, the eddy vertical PV flux is normalized using the background PV value. If the eddy vertical PV flux is sufficiently large compared to the background PV value, the vertical diffusion coefficient becomes large.

3. Mean meridional circulation in the extratropical tropopause region

a. Each wave group’s E-P flux

Figures 1a–c show the E-P flux and its divergence for each of the three wave groups for January, together with the 2- and 4-PVU (1 PVU = 10^{-6} K m^2 kg^{-1} s^{-1}) surfaces; these PVU surfaces approximately correspond to the dynamical tropopause in the extratropics. Although the 2-PVU surface has widely been used to define the dynamical tropopause, it has been known to lie close to the bottom of the ExTL (e.g., Hoor et al. 2004) rather than the thermal tropopause [defined by the lapse rate profile following WMO (1957)]. Part I showed that the 4-PVU surface is closer to the thermal tropopause than the 2-PVU surface in the high-resolution GCM.

Both the PW and MW E-P fluxes propagate upward in the troposphere and strongly converge in the upper troposphere and around the tropopause in the extratropics. The convergence of these E-P fluxes decelerates westerly wind. The convergence is stronger in the winter hemisphere than in the summer hemisphere. In the extratropical tropopause region, especially below \( \theta = 325 \) K, the PW E-P flux converges more obviously than the MW E-P flux. From the subtropics to midlatitudes (latitudes of about 30°–40°), both the PW and MW E-P fluxes diverge in the lowermost stratosphere.

In contrast to the PW and MW E-P fluxes, the GW E-P flux diverges around and just above the tropopause from the subtropics to high latitudes, accelerating the westerly wind. This acceleration is commonly seen in both hemispheres. The strongest GW E-P flux divergence centers on the 40°–60° latitudes near the tropopause. The GW EP-flux divergence in the extratropical tropopause region may be affected by the spontaneous generation of inertia–gravity waves from jets and fronts. O’Sullivan and Dunkerton (1995) revealed the excitation and propagation of gravity waves near the tropopause related to baroclinic waves. Plougonven and Snyder (2007) also indicated a strong gravity wave packet in the tropopause region in the baroclinic life cycle dominated by cyclonic behavior.

The PW E-P flux obviously converges in the extratropical overworld stratosphere. The breaking of PWs leads to a strong deceleration of the westerly wind. This explains most of the deceleration by the total wave
components. In the subtropical stratosphere, the convergence of the GW E-P flux largely influences the zonal mean zonal wind. This is more substantial in the winter hemisphere than in the summer hemisphere.

b. Wave-driven mean meridional circulation

The downward control calculation using the E-P flux demonstrates that the PW group explains most of the forcing that drives the mean meridional circulation in the extratropical tropopause region and stratosphere (Figs. 1d–f). The convergence of the PW E-P flux drives a poleward-downward Brewer–Dobson circulation in the extratropical stratosphere and a fast mean poleward flow near and just below the extratropical tropopause. These velocities are much larger in the winter hemisphere than in the summer hemisphere. The MW group also induces a poleward and downward circulation in the extratropical upper troposphere. These facts have been commonly suggested by previous studies using coarse-resolution GCMs (e.g., Rosenlof and Holton 1993).

An analysis of the high-resolution GCM shows that the GW group causes a mean equatorward flow in the extratropical tropopause region (latitudes of 25°–70°, below approximately \( \theta = 350 \) K), which is associated with its E-P flux divergence. This differs from the mean poleward flow induced by the PW and MW groups. The

![Figure 1: Latitude–potential temperature cross section of monthly mean](image)
The GW group also drives a mean poleward flow in the subtropical lower stratosphere (above approximately $5350\text{ K}$) and a mean downward flow in the midlatitude lower stratosphere (latitudes of about $40^\circ$–$50^\circ$). These flows are stronger in the winter hemisphere than in the summer hemisphere, corresponding to an interhemispheric difference in the GW E-P flux convergence. The poleward and downward circulation acts to transport air from the subtropical stratosphere to the midlatitude tropopause region.

It should be noted that in addition to the E-P flux divergence, the time tendency of the zonal mean zonal wind influences the mean meridional circulation in the downward control calculation. However, this is not included in the analysis [Eq. (1)], which causes some disagreements in the mean meridional circulation between the total wave components’ downward control calculation and the residual mean meridional circulation around the subtropical jet core.

4. Mixing processes in the extratropical tropopause region

a. Analysis of diffusion coefficients

1) Meridional structure in isentropic coordinates

Figure 2 depicts the meridional cross section of the logarithms of the (a),(b) isentropic diffusion coefficient ($K_{yy}, \text{ m}^2\text{s}^{-1}$) and (c),(d) vertical diffusion coefficient ($\mathcal{Q}_{q}/q, \text{ K s}^{-1}$) for (a),(c) January and (b),(d) July. The black contour lines represent the zonal mean zonal wind with intervals of $15\text{ m s}^{-1}$. The white contour lines represent the absolute values of PVU at 2 and 4 PVU.
reveal obvious isentropic mixing across the tropopause below approximately $\theta = 340$ K during winter at the poleward side of the subtropical jet stream. Active cross-tropopause mixing below $\theta = 340$ K has been commonly revealed by Lagrangian trajectory calculations (Berthet et al. 2007) and the relationships between different chemical species obtained from satellite measurements (Hegglin et al. 2009). The average level of 340–350 K for the ExTL top found by Hegglin et al. (2009) suggests that the troposphere-to-stratosphere transport (i.e., mixing) is very active below this level. Birner (2006) evaluated the approximated observed PV structure of the TIL and found weak PV gradients confined to the TIL; these are suggestive of active large-scale dynamical mixing within the TIL. These observational analyses also appear to fit the high-vertical-resolution GCM analysis.

In the stratosphere ($\theta > 380$ K; i.e., the overworld stratosphere) of the winter hemisphere, the strong isentropic mixing revealed by the large $K_{yy}$ values coincides with weak westerly wind. In comparison to that in the stratosphere, $K_{yy}$ is smaller in the lowermost stratosphere (between approximately $\theta = 330$ and 380 K), implying less breaking of atmospheric waves in the middleworld stratosphere. The value of $K_{yy}$ is smaller in the summer hemisphere than in the winter hemisphere throughout the extratropical UTLS. A very small $K_{yy}$ value is observed over the Antarctic throughout the stratosphere in all seasons.

The value of $K_{yy}$ is very small around the subtropical jet core between approximately $\theta = 330$ and 390 K where the atmospheric waves break less, likely because of very strong westerly winds. The weak mixing suppresses the air exchange between the tropical upper troposphere and the extratropical lower stratosphere across the subtropical tropopause. The $K_{yy}$ values around the subtropical jet core are much smaller in the winter hemisphere than in the summer hemisphere, corresponding to a seasonal difference in zonal mean westerly wind speed as well as monsoon circulation activity in the Northern Hemisphere. The stronger isentropic mixing around the subtropical tropopause is considered to make the chemical character of the extratropical lowermost stratosphere more tropospheric during summer than in other seasons (e.g., Hoor et al. 2004). A locally enhanced meridional PV gradient at the jet core may also inhibit Rossby wave breaking and cause minima in $K_{yy}$ in the jet (Hitchman and Huesmann 2007). The value of $K_{yy}$ is larger just below the subtropical jet stream than around the jet core.

These general characteristics of isentropic mixing derived from the $K_{yy}$ using the high-resolution GCM output have been commonly revealed by $K_{yy}$ estimates (Newman et al. 1988; Bratseth 1998) and the effective diffusivity (Haynes and Schuckburgh 2000) using the output of coarse-resolution models. The analysis of the high-vertical-resolution GCM indicates a larger maximum value and larger vertical variations of $K_{yy}$ around the extratropical tropopause than those produced by coarse-resolution models.

Vertical mixing, on the other hand, is enhanced in the extratropical tropopause region at the midlatitudes (latitudes of approximately 30°–60°), with larger values and areas in the summer hemisphere than in the winter hemisphere (Figs. 2c,d). Large vertical diffusion coefficient values are located between approximately 1 and 8 PVU. This region is almost collocated with a mixing layer associated with stirring and mixing in the tropopause region identified from aircraft observations of ozone and carbon monoxide concentrations (Brioude et al. 2008). Hegglin et al. (2005) suggested that mixing across isentropes (i.e., vertical mixing) is more important than isentropic mixing in transporting chemical constituents in the lowermost stratosphere. In contrast, a very small vertical diffusion coefficient value is present just above the mixing layer. This small value suggests the existence of a vertical mixing barrier between the ExTL and higher levels. This vertical mixing barrier is associated with a very weak diabatic heating–cooling rate at this level as a consequence of the cancellation between the shortwave heating by ozone and longwave cooling by water vapor (cf. Fig. 14b in Part I).

2) MERIDIONAL STRUCTURE IN TROPOPAUSE-BASED COORDINATES

Figure 3 shows the diffusion coefficients in tropopause-based coordinates. The tropopause-based coordinates exhibit a more compact structure for the mixing layer relative to the tropopause, as compared to the standard coordinates in which the time average is taken at a constant height (or pressure) level. We first analyzed $K_{xy}$ in isentropic coordinates and then converted it into tropopause-based coordinates. In other words, the temporal average of the analyzed $K_{xy}$ was taken at a constant distance with respect to the zonal mean tropopause height (determined by the isentropic zonal mean temperature profile). Although this calculation method does not appropriately reflect local atmospheric structures related to longitudinal tropopause variations, it is still useful for demonstrating meridional transport processes in association with the mean tropopause location. On the other hand, $K_{yy}$ as analyzed directly in standard tropopause-based coordinates (where the zonal average is taken at a constant distance with respect to the local tropopause height) cannot be physically interpreted (i.e., $K_{yy}$ does not represent mixing by adiabatic wave motions; see also discussions in Part I).
The value of $K_{yy}$ is large between approximately $\Delta \theta_{TP} = -20$ K ($-2.5$ km) and $+10$ K ($+1$ km) throughout the extratropics of the winter hemisphere. Vertical mixing is very active at the midlatitudes around the tropopause above approximately $\Delta \theta_{TP} = -20$ K, especially in the summer hemisphere, but is substantially suppressed at $\Delta \theta_{TP} = +10$ to $+15$ K throughout the extratropics. Consequently, both the isentropic and vertical components of the eddy motions strongly diffuse air within the extratropical tropopause region between approximately $\Delta \theta_{TP} = -20$ and $+10$ K.

The thickness of the simulated mixing layer is very consistent with that of the observed ExTL. Hegglin et al. (2009) identified an annual mean thickness of the ExTL to be approximately about 3 km in the NH extratropics, using satellite observations of chemical species and their correlations. Hoor et al. (2004) found an ExTL depth of about 20–25 K in winter and 30 K in summer. The similarity in the depths of the simulated mixing layer and observed ExTL implies that the mixing processes have considerable influences on the location and depth of the ExTL. It should be noted that mean downward motions, together with mixing processes, distinctly affect chemical tracer concentration gradients (i.e., depth and seasonality of the ExTL), especially at the lower part of the ExTL (Part I).

3) SEASONAL VARIATION IN TROPOPAUSE-BASED COORDINATES

The diffusion coefficients in tropopause-based coordinates show clear seasonal variations in the extratropical UTLS, as depicted in Figs. 4 and 5. A large $K_{yy}$ value is observed in the extratropical tropopause region throughout the year between approximately $\Delta \theta_{TP} = -20$ and $+10$ K at 60°N, with a maximum in December and a minimum in August (Fig. 4a). The thickness of the layer with strong isentropic mixing is larger during winter than during summer. The upper boundary level of the
FIG. 4. Seasonal variation of (a),(c) the logarithms of the isentropic diffusion coefficient ($K_{yy}$, m$^2$ s$^{-1}$) and (b) the vertical diffusion coefficient ($Q'q'/q$, K s$^{-1}$) at (a),(b) 60°N and (c) 30°N, as a function of the distance from the thermal tropopause ($\Delta \delta_{TP}$). The black contour lines represent the zonal mean zonal wind with intervals of 5 m s$^{-1}$. The white contour lines represent the absolute values of PVU at 2 and 4 PVU.
mixing layer remains almost constant throughout the year, whereas the lower boundary level is about 10 K lower in winter than in summer. A large $K_{yy}$ at the tropopause level indicates strong mixing between the upper troposphere and lowermost stratosphere along isentropes. Strong isentropic mixing at the tropopause level is present at latitudes of approximately 40$^\circ$–80$^\circ$ in both hemispheres, reaching a maximum at around 50$^\circ$N and 60$^\circ$S in midwinter (Fig. 5a). The seasonal and latitudinal variations in $K_{yy}$ at the tropopause level appear to correspond well with those in baroclinic wave activity, implying that baroclinic waves act to effectively flatten chemical tracer concentration gradients along isentropes at the tropopause level at these latitudes.

In the overworld stratosphere (above $\Delta \theta_{TP} = +60$ K) at 60$^\circ$N, $K_{yy}$ is large from November to February and very small from June to September (Fig. 4a). Compared to that in the overworld stratosphere, $K_{yy}$ has less seasonality in the middleworld stratosphere (between approximately $\Delta \theta_{TP} = +30$ and +60 K). Also, $K_{yy}$ and zonal mean zonal wind show similar seasonal variations in the stratosphere.

Vertical mixing in the extratropical tropopause region is very strong from June to August and weak from November to March at 60$^\circ$N (Fig. 4b). This seasonal
variation is very different from that of isentropic mixing. Strong vertical mixing across the tropopause occurs at latitudes of 40°–65° during summer in both hemispheres (Fig. 5b). The summertime active vertical mixing at midlatitudes may be strongly associated with three-dimensional turbulence and convection. It appears that the active vertical mixing results in smaller CO concentration gradients across the tropopause during summer than in other seasons (Hoor et al. 2004; Hegglin et al. 2009). In contrast, the vertical mixing suppression related to a very weak diabatic heating rate persists throughout the year at around ΔΘTP = +10 to +20 K.

Around the subtropical tropopause, both Kyy and zonal mean zonal wind show obvious seasonal variations (Fig. 4c). A small Kyy is related to a strong westerly wind from November to April, when atmospheric waves do not break easily. In contrast, from June to August, Kyy is large between approximately ΔΘTP = –20 to +40 K, allowing mixing between the tropical upper troposphere and extratropical lower stratosphere, as consistently reported by observational studies (e.g., Hoor et al. 2004). The strong isentropic mixing extends across the latitudes of 15°–35° in both hemispheres (Fig. 5a).

b. Implications from 2D model simulations

In section 4a, we demonstrated the characteristics of the mixing processes. In this section, we briefly demonstrate the roles of these mixing processes in determining the PV (or constituent) distributions. For this purpose, we compare the PV distributions of the GCM simulation and the integration results of a zonal mean PV continuity equation based on mass-weighted isentropic zonal means (MIM) analysis [Part I; i.e., a two-dimensional (2D) model simulation]:

\[
\frac{\partial \tilde{q}^*}{\partial t} = -\frac{\tilde{v}^*}{a} \frac{\partial \tilde{q}^*}{\partial \phi} - \frac{\tilde{\theta}^*}{a \cos \phi} \frac{\partial \tilde{q}^*}{\partial \theta} - \frac{1}{a \cos \phi} \left( \frac{\partial (\nu \tilde{q}^*)}{\partial \phi} \cos \phi \right) \\
- \frac{1}{\rho_0} \frac{\partial \rho_0 (\theta' q')^*}{\partial z} + \left( \frac{\partial \tilde{\theta}^*}{\partial \theta} \right),
\]

(6)

The first term on the right-hand side of Eq. (6) represents the PV time tendency due to mean meridional transport; the second term is that due to mean vertical transport; the third is that due to eddy meridional transport; the fourth is that due to eddy vertical transport, and the fifth is that due to source–sink-related changes in stability through diabatic processes. Although the PV structure is associated strongly with the atmospheric stability profile, its temporal evolution is caused by atmospheric transport processes together with diabatic source–sink effects. By excluding the eddy transport terms in the zonal mean equation and comparing the simulation result with the GCM simulation, we show the effects of the mixing processes.

In a standard 2D model simulation, we neglected the influence of the eddy meridional and vertical transport terms in the zonal mean equation [Eq. (6)] but considered the PV tendencies using the mean transport and nonconservative (diabatic source–sink) terms. A comparison between this 2D model simulation and the GCM simulation result demonstrates the importance of the eddy transport. In addition, another 2D model simulation was performed that considered the eddy vertical transport term together with the mean transport and nonconservative terms (i.e., only the eddy meridional transport term was neglected), demonstrating the importance of isentropic mixing on PV distributions. The mean meridional circulation fields and static stability profiles obtained from the GCM simulation results were provided for the 2D model simulation to calculate the mean transport and nonconservative terms. Twenty-day integrations were performed starting at 1 January and 1 July, using the initial PV distributions obtained from the GCM simulation results.

Figure 6 compares the zonal mean PV profiles between the GCM and 2D model simulations averaged over 50°–60°N for 14–20 January and 14–20 July. The 2D model that neglected both the eddy meridional and vertical transport terms has a larger PV value around the tropopause and a much smaller vertical PV gradient just above the tropopause compared to the GCM during winter. As a result of excluding the eddy transport terms, the maximum PV gradient height falls to about 10 K near the tropopause. The lower height of the maximum gradient is commonly revealed by another 2D model simulation that includes the eddy vertical transport. The active isentropic mixing identified from a large Kyy strongly diffuses air within the extratropical tropopause region between approximately ΘTP = –20 and +10 K, lifting the maximum PV gradient from just below to just above the tropopause. This effect makes an important contribution to the formation of large tracer gradients in the upper part of the ExTL.

During summer, the maximum PV gradient simulated using the GCM is located near ΔΘTP = +5 K. The 2D model that neglects both the eddy meridional and vertical transport terms indicates a lower height for the maximum PV gradient around the tropopause. By including the eddy vertical transport term, the maximum PV gradient height and PV profile in the 2D model simulation approach those in the GCM just above the tropopause. Very strong vertical mixing occurs between approximately ΘTP = –20 and +10 K at the NH midlatitudes during summer. Variations in vertical mixing (i.e., convergence of eddy vertical fluxes) act to shift the
vertical PV gradient from around to just above the tropopause. The results imply that vertical mixing is more important for determining the height of the large PV (or constituent) gradient around the tropopause during summer than during winter, in association with the strong convergence of eddy vertical fluxes at just above the tropopause (i.e., the vertical mixing barrier).

c. Wave characteristics: Isentropic mixing

1) Vertical profile

To investigate the relative contributions of atmospheric motions with different horizontal scales to isentropic mixing, the eddy PV flux can be decomposed into individual zonal wavenumber components using FFT. The eddy PV fluxes at the different zonal wavenumbers \( s \) normalized by the background PV gradient measure the contributions of the different horizontal-scale atmospheric disturbances to the total \( K_{yy} \):

\[
K_{yy}(s) = \left( \frac{q^2 \frac{\partial q}{\partial \theta}}{a \frac{\partial \theta}{\partial \phi}} \right)_{\theta}
\]

Figures 7 and 8 depict the zonal wavenumber dependence of the normalized isentropic eddy PV flux averaged...
over 50°–60°N (where mixing in both the isentropic and vertical directions is very strong around the tropopause) in January. The isentropic eddy PV flux is strongly influenced by large and medium horizontal-scale motions, particularly those with $s \leq 12$–13 below $\theta = 325$ K and $s \leq 5$–6 above the level. In other words, planetary waves largely contribute to isentropic mixing in the stratosphere, while both planetary-scale and synoptic-scale waves cause strong isentropic mixing below the lowermost stratosphere. Medium-scale waves with horizontal wavelengths of a few thousand kilometers (Sato et al. 1993) also seem to contribute to the isentropic mixing below the lowermost stratosphere. Small $K_{xy}$ values in the middleworld stratosphere (cf. Fig. 2a) mostly arise from small contributions by large horizontal-scale motions, especially around $\theta = 370$–400 K.

**FIG. 7.** Zonal wavenumber–potential temperature cross section of isentropic eddy PV flux normalized by background PV gradient (m$^2$ s$^{-1}$) averaged between 50° and 60°N in January. The straight lines represent the absolute values of PVU at 2 and 4 PVU. Note the different scaling used in the different panels of the figure. The right panels show the monthly mean $N^2$ profiles ($10^{-4}$ s$^{-2}$).

**FIG. 8.** Relative contributions of different wave components to isentropic eddy PV flux normalized by background PV gradient (left) in m$^2$ s$^{-1}$ and (right) in % averaged between 50° and 60°N in January. These are associated with the zonal wavenumbers 1–3 (thin solid line), 4–7 (dotted line), 8–20 (dashed line), and 21–213 (bold solid line). The straight lines represent the PVU values at 2 and 4 PVU.
It is worth noting that atmospheric motions with horizontal wavelengths of less than a few thousand kilometers play essential roles in isentropic mixing just above the extratropical tropopause (as seen in Fig. 8b). These small horizontal-scale motions (i.e., \( s \geq 21 \)) cause stronger isentropic mixing between \( \theta = 330 \) and 355 K than at other levels in the stratosphere. Both the small horizontal-scale motions and static stability are maximized at around 340–350 K (Fig. 7), suggesting that active isentropic mixing is connected with rapid changes in the static stability through small-scale dynamics around the TIL.

Figure 8 compares the relative contribution of each wave group to the total isentropic mixing at the NH mid-latitudes in January. The small horizontal-scale motions (i.e., \( s \geq 21 \)) provide an important contribution to the total isentropic mixing just above the tropopause, with a contribution of about 20\%. Although large horizontal-scale motions dominate the strength of the isentropic mixing in the extratropical UTLS (the contribution of planetary wave–scale motions (i.e., \( s \leq 3 \)) is more than 30\% in the troposphere and reaches 45\% at 450 K), the enhanced contribution of the small horizontal-scale mixing is a unique and important characteristic of the extratropical tropopause region.

2) SEASONAL AND INTERHEMISPHERIC DIFFERENCES

The dominant horizontal scales appearing in the isentropic mixing differ somewhat in the two hemispheres and in different seasons in the extratropical UTLS, as depicted in Fig. 9. At the NH midlatitudes during summer, large horizontal-scale mixing is the strongest just below the tropopause and weakens rapidly with height in the stratosphere (Fig. 9a), which is different from winter. In particular, large horizontal-scale mixing, with \( s = 1–4 \), makes a much weaker contribution to the total isentropic mixing in the stratosphere during summer than during winter. The seasonal variation in the planetary wave activity dominates that of the \( K_{yy} \) in the NH stratosphere.

Isentropic mixing, with \( s = 1–2 \), is much weaker in the SH than in the NH during winter (Fig. 9b). Quasi-stationary planetary wave activity is weaker in the SH than in the NH. This is responsible for the weaker large-scale mixing in the SH. In contrast, synoptic-scale mixing, with \( s = 4–7 \), is somewhat stronger in the SH than in the NH in the upper troposphere and around the tropopause, which is attributable to the interhemispheric difference in baroclinic wave activity.

Small horizontal-scale mixing just above the extratropical tropopause is stronger in winter than in summer, and in the NH than in the SH. In the SH, small horizontal-scale mixing is strongest just below the tropopause, just as in the NH, but rapidly weakens with height in the lower stratosphere, which is different from the NH (Fig. 9b).

In the subtropical UTLS during summer, isentropic mixing is dominated by large horizontal-scale mixing, with \( s = 1–7 \) (Fig. 9c). This suggests that the active cross-tropopause mixing is attributable to Rossby wave breaking in the subtropics during summer, as commonly suggested by Postel and Hitchman (1999). Small horizontal-scale mixing reaches a maximum in the lower stratosphere between 390 and 420 K in the subtropics, collocating with the strong GW E-P flux convergence (cf. Fig. 1c).

3) WAVE CHARACTERISTICS: VERTICAL MIXING

Here we consider the occurrence of vertical mixing due to diabatic processes around the TIL. The vertical mixing processes around the TIL seem to obviously change the vertical gradients of the chemical constituent concentrations around the ExTL. To review the vertical mixing characteristics in the TIL, Figs. 10 and 11 show the vertical profile of the eddy vertical PV flux normalized by the background PV value for each zonal wavenumber (which is proportional to the vertical diffusion coefficient [Eq. (5)]:

\[
\frac{\partial q}{\partial z} = \sqrt{Q_{yy} q_{yy}} \quad (8)
\]

The vertical dispersion by small-scale eddy motions with horizontal wavelengths of less than a few thousand kilometers is clearly strengthened just above the tropopause between approximately \( \theta = 330 \) and 360 K (Fig. 11), whereas that by large-scale eddy motions is weaker at this level compared to those at other levels. Small horizontal-scale motions, with \( s \geq 21 \), provide dominant contributions to the total vertical mixing in the lower stratosphere below 400 K (Fig. 12), reaching a maximum between 345 and 375 K.

As a result, small horizontal-scale motions lead to strong eddy PV fluxes in both the isentropic and vertical directions just above the tropopause (cf. Figs. 8 and 12). The analysis results confirm the occurrence of obvious three-dimensional mixing by small-scale motions around the TIL. It should be noted that isentropic coordinates might not be suitable for representing vertical mixing related to overturning isentropes involved in breaking gravity waves.

4) SPATIAL DISTRIBUTION

Figure 12a shows the spatial distribution of the monthly mean amplitude of small horizontal-scale PV disturbances for \( s \geq 21 \) on the 360-K surface located just above
FIG. 9. As in Fig. 7, but averaged between (a) 50°N and 60°N in July, (b) 50°S and 60°S in July, and (c) 28° and 32°N in July. Note the different scaling used in the different panels of the figure.
the tropopause in the extratropics (i.e., in the TIL) but in the upper troposphere in the tropics. The small horizontal-scale PV disturbances, probably strongly associated with small-scale variability in $N^2$, are large over strong convection regions (i.e., the maritime continental region), mountainous regions (i.e., over the Rocky Mountains, the Himalayas, Iceland, and the Antarctic Peninsula), and the polar sides of the subtropical westerly jet stream. A snapshot picture (Fig. 12b) clearly exhibits the horizontal structures (including the lateral propagations) of the small horizontal-scale PV disturbance over strong convective regions (e.g., the southeast Asian Maritime Continent), frontal systems (e.g., in the SH east Pacific), and mountainous regions. These small-scale disturbances occur over possible sources of gravity waves (i.e., high mountains, cyclones, fronts, and convection), suggesting that the propagation and breaking of gravity waves lead to active small-scale mixing in the TIL. Three-dimensional gravity wave propagation in the high-resolution GCM is further discussed in Watanabe et al. (2008) and Sato et al. (2009).

![Figure 10](image1.png)

**Fig. 10.** As in Fig. 7, but for the eddy vertical PV flux normalized by the background PV value (K s$^{-1}$) averaged between 50° and 60°N in January. Note the different scaling used in the different panels of the figure.

![Figure 11](image2.png)

**Fig. 11.** As in Fig. 8, but for the relative contributions of the different wave components to the vertical eddy PV flux normalized by the background PV value averaged between 50° and 60°N in January.
5. Discussions: Small-scale mixing around the TIL

In this section, we discuss possible mechanisms for the three-dimensional mixing processes (i.e., isentropic and vertical mixing) in the TIL associated with atmospheric waves. Hegglin et al. (2009) proposed that Rossby wave breaking events cause filaments of tropospheric air advected into the lowermost stratosphere, and that these filaments may be three-dimensionally mixed by small-scale motions (i.e., turbulence generated by dynamical processes such as gravity waves). Gravity wave dissipation occurs when the waves break or encounter critical
levels. Meanwhile, even if gravity waves do not break or encounter critical levels, large-amplitude gravity waves can have eddy vertical dispersions and induce net three-dimensional mixing under nonuniform diabatic heating rate fields. In other words, zonal variations in the diabatic heating rate on wave (or isentropic) surfaces generate the eddy components of the vertical dispersions away from the surfaces.

First, we investigate the gravity wave energy around the TIL. Figure 13 depicts the potential energy $PE = \frac{1}{2} \left( \frac{g^2}{N^2} \right) (\theta'^2 + \theta''^2)$ and kinetic energy $KE = \frac{1}{2} (u'^2 + v'^2 + w'^2)$ of gravity waves in January. Small horizontal-scale fluctuations with total wavenumbers $n$ greater than 22 are analyzed as gravity waves, although filamentary structures also contribute to the small-scale fluctuations around the tropopause (figure not shown). The large total energy ($TE$) of the gravity waves ($TE = KE + PE$) is observed from the upper troposphere to just above the tropopause in the extratropics, which is mostly attributable to $KE$. A local maximum $PE$ also contributes to the large $TE$ just above the tropopause. As a result, the ratio of $PE$ to $KE$ ($PE/KE$; Fig. 13d) is maximized in the middle troposphere and just above the tropopause and is minimized just below the tropopause. A local maximum for the relative ratio lying just above the tropopause coincides with the $N^2$ maximum. The enhancement of gravity wave energy near the tropopause has been commonly investigated by radiosonde and GPS radio occultation data (e.g., Allen and Vincent 1995; Schmidt et al. 2008) and is considered to be an important characteristic of the extratropical tropopause region.

Second, we consider the dissipation and saturation of the gravity waves around the TIL. Following the linear wave theory (e.g., Lindzen 1981), the breaking level of gravity waves is defined as the level where the isentrope first becomes $(\partial \theta / \partial z) = (\partial \theta / \partial z) + (\partial \theta / \partial z) = 0$. The assumption in this theory is that any wave amplitude in excess of the threshold value causes instability and turbulence and acts to prevent further growth of the wave amplitude (i.e., the waves become convectively unstable).
In the TIL, the zonal mean static stability is large, making the gravity wave saturation amplitude large. The large-amplitude gravity waves can cause vertical eddy dispersions owing to nonuniform diabatic fields. In addition, the large-amplitude gravity waves may be easily dissipated because of radiative relaxation. In contrast, in the upper part of and just above the TIL, as the static stability decreases with height (and the gravity wave amplitude increases with height because of the exponential decrease of atmospheric density), gravity waves may easily reach their saturation limit and breaking level and cause three-dimensional turbulence and mixing.

Meanwhile, dynamical instability plays a significant role in the saturation of gravity waves. Gravity wave saturation is significant in the enhancement of momentum flux deposition and turbulent diffusion (e.g., Dunkerton 1989). The dynamical instability occurs in the presence of a vertical mean wind shear (e.g., around the tropopause owing to the presence of a jet stream). To measure the stability of the gravity waves, we can estimate the probability of the occurrence of the dynamical instability based on the Richardson number (Ri):

$$\text{Ri} = \frac{N^2}{\left(\frac{\partial u}{\partial z}\right)^2}. \quad (9)$$

The atmosphere is dynamically unstable when $\text{Ri} < 0.25$. Figure 14 shows that this dynamical instability occurs more frequently in the extratropical tropopause region between approximately $\Delta \theta_{TP} = -10$ K and $+30$ K, compared to other levels, where large-amplitude inertia–gravity waves may undergo breaking. The region of large dynamical instability almost corresponds to the strong small-scale three-dimensional mixing region (cf. section 4).

Consequently, these small-scale dynamics in association with the dissipation and saturation of gravity waves seem to explain the obvious three-dimensional mixing by small-scale motions around the TIL and might be the essential cause of the large constituent concentration gradients around the ExTL. However, further study is still required to comprehend gravity wave processes related to the rapid change in the static stability around the TIL (e.g., generation mechanisms, changes in vertical wavelength).

### 6. Conclusions

The roles of atmospheric processes with various scales, including gravity wave–scale to planetary-scale phenomena, in the transport and mixing processes in the extratropical tropopause were examined using a high-vertical-resolution GCM. The GCM has a vertical resolution of about 300 m above the extratropical upper troposphere and represents the dynamic and thermal finescale structure of the extratropical tropopause region. The GCM made it possible to resolve most of the gravity waves and represent their propagation and breaking in the lower and middle atmosphere. Our analysis of atmospheric processes with different scales provides a comprehensive view of transport and mixing in the extratropical tropopause region (Fig. 15). The dominant transport scale differs between the tropopause region and other levels in the UTLS.

The relative contributions of atmospheric waves with different scales, including resolved gravity waves, to the driving of the mean meridional circulation have been examined using the downward control calculation. The E-P flux associated with gravity waves diverges and induces a mean equatorward flow in the extratropical tropopause region. The mean equatorward flow induced by the gravity wave E-P flux divergence partly cancels a mean poleward flow induced by the planetary and synoptic-scale wave E-P flux convergence near the extratropical tropopause. The gravity waves also drive a mean poleward circulation in the subtropics and a mean downward circulation at the midlatitudes in the lower stratosphere of the winter hemisphere.
The diffusion coefficients estimated from the eddy PV fluxes in tropopause-based coordinates were used to investigate the significance of the eddy transport in the extratropical tropopause region. The air within the extratropical tropopause region between approximately $D_{TP} = 52$ K and 110 K was found to be strongly diffused due to adiabatic eddy motions from late autumn to early spring. Vertical eddy motions are substantial in the tropopause region at the midlatitudes during summer but are strongly suppressed at $D_{TP} = 10$ to 15 K just above the well-mixed region throughout the year. The simulated mixing layer almost coincided with the observed ExTL, suggesting that mixing processes play important role in determining the depth and location of the ExTL. The value of $K_{yy}$ was found to be very small around the subtropical westerly jet, inhibiting the transport of tropical tropospheric air into the extratropical lowermost stratosphere.

It should be emphasized that small-scale motions with horizontal wavelengths of less than a few thousand kilometers were found to cause obvious three-dimensional mixing just above the extratropical tropopause (i.e., in the TIL), as depicted in Fig. 15b. This small horizontal-scale mixing in the TIL is stronger during winter than during summer and in the NH than in the SH. In contrast, large-scale waves cause strong isentropic mixing below the lowermost stratosphere. Small horizontal-scale disturbances are large over regions with possible sources of gravity waves (high mountains, cyclones, fronts, and convection) at the TIL level. The analysis of the gravity wave energy proved an enhancement of gravity wave PE just above the tropopause. The enhanced PE contributes to the increase in total gravity wave energy in the TIL, along with a large KE.

In the TIL, the saturation amplitude of the gravity waves can be large because of a large static stability. This situation allows the occurrence of large-amplitude gravity waves. Even when gravity waves do not break, large-amplitude gravity waves can be dissipated because of radiative relaxation and can cause eddy vertical dispersions under nonuniform diabatic heating rate fields. In contrast, in the upper part of and just above the TIL, gravity waves may easily reach their saturation limit and breaking level because of a decrease in static stability and atmospheric density with height. Dynamic instability occurs more frequently in the extratropical tropopause region than at other levels, which also may lead to the saturation of gravity waves. These situations associated with the dissipation and saturation of gravity waves seem to cause small-scale three-dimensional mixing around the TIL.

To summarize, the small-scale dynamics associated with the propagation and breaking of gravity waves plays important roles in driving tracer transports by both the mean meridional circulation and the three-dimensional mixing in the extratropical tropopause region. It may be
important to include these small-scale dynamic effects in GCMs or CTMs to obtain better simulations of the TIL and ExTL.

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