Effective diffusivity as a diagnostic of atmospheric transport

2. Troposphere and lower stratosphere

Peter Haynes and Emily Shuckburgh
Centre for Atmospheric Science, Department of Applied Mathematics and Theoretical Physics, Cambridge, England

Abstract. The effective diffusivity diagnostic is used to analyze the isentropic transport and mixing properties of observed winds in the upper troposphere and the lower stratosphere (300–450 K), following the approach described in part 1 [Haynes and Shuckburgh, this issue]. Local minima in effective diffusivity on isentropic surfaces in the range 330–400 K indicate transport barriers in each hemisphere associated with the extratropical tropopause. The strongest part of these “tropopause barriers” are coincident with the core of the subtropical jet at about 350 K. They are shown to have a seasonal evolution in which they are strongest in winter and considerably weakened by the monsoon circulations in summer. The barrier in the Southern Hemisphere is seen to be generally stronger than that in the Northern Hemisphere during the same season. The minimum value of effective diffusivity on each isentropic surface is proposed as a new definition of the tropopause. This effective-diffusivity definition corresponds most closely to potential vorticity (PV) values of ±2 PVU at 330 K, ±2.5 PVU at 350 K, and ±4.5 PVU at 370 K (with larger values being more appropriate during the summer monsoon period), rather than to the conventional tropopause definition of a single PV value at all levels. It is also demonstrated that the lower limit of the barrier at the stratospheric polar-vortex edge, i.e., the “sub-vortex” transition, varies in altitude throughout the winter. In the Antarctic the transition generally occurs at 380 K and is sometimes as low as 350 K. In the Arctic the transition is higher, rarely occurring below 400 K and frequently occurring above 450 K.

1. Introduction

In part 1 [Haynes and Shuckburgh, this issue] the effective diffusivity diagnostic [Nakamura, 1996] was used to identify successfully the barriers to isentropic eddy transport in the stratosphere (which are shown in schematic Figure 1, part 1), to quantify their relative strengths, and to analyze how they vary with altitude and through the seasons. The method used was to advect a passive “test tracer” using observed isentropic winds and to calculate the effective diffusivity from the instantaneous tracer field as it evolves. The effective diffusivity can be regarded as a measure of the complexity of the geometric structure of the tracer. In regions of strong mixing the geometric structure will be complex, and the effective diffusivity large; in barrier regions the geometric structure will be simple, and the effective diffusivity small.

The results presented in part 1 showed clearly the division of the stratosphere by eddy transport barriers at the polar vortex edge and in the subtropics into the distinct regions suggested by observed inhomogeneous chemical distributions. Inhomogeneous chemical distributions are also observed in other parts of the atmosphere, for example the range of “middleworld” isentropic surfaces (approximately 320–380 K) which lie partly in the stratosphere and partly in the troposphere, and consequently the identification and quantification of barriers to isentropic eddy transport is not only relevant to the stratosphere. The strong contrasts in concentrations of chemical species, such as ozone and water vapor, between the tropopause and the stratosphere strongly suggest that the tropopause is also a barrier to transport (as indicated in Figure 1, part 1), and therefore that rates of stratosphere-troposphere exchange may be determined in part by the leakage rate across this barrier [e.g., Holton et al., 1995]. Stratosphere-troposphere exchange along the middleworld isentropic surfaces has been studied by following the evolution of ensembles of particles [Pierrehumbert and Yang, 1993], of a tracer field [Chen, 1995], and of material contours [Appenzeller et al., 1996a; Bithell and Gray, 1997]. However, those studies which have attempted to identify objectively the tropopause as a barrier to transport by calculating stretching rates [Pierrehumbert and Yang, 1993, Bithell and Gray, 1997] have conspicuously failed.

Compared to the vortex-edge barrier, the tropopause barrier is rather permeable: if particles are placed in the barrier region initially, they may leave that region after only a limited time. Thus stretching rates calculated on the basis of evolution over longer times may not detect the smaller
stretched rates associated with the barrier region. In particular, an approach based on particles or material contours may be expected to fail completely when the timescale for a particle to leave a barrier region is less than the timescale necessary to define a stretching rate. It seems that effective diffusivity, being a hybrid Eulerian-Lagrangian diagnostic as noted in part 1, section 2.1, may overcome some of these difficulties.

Hence, here in part 2, we apply the approach based on effective diffusivity introduced in part 1, to investigate isentropic transport and mixing in the lower stratosphere and the troposphere. We begin in section 2 by discussing and demonstrating the validity of applying our method to this part of the atmosphere.

In section 3 we show that a tropopause transport barrier can indeed be identified as a minimum in effective diffusivity on isentropic surfaces in the middle-latitude region.

The only previous study to have systematically investigated the vertical variation in strength of isentropic stratosphere-troposphere exchange is Chen [1995], who considered advection of a passive tracer by observed winds on isentropic surfaces 310–360 K. Chen distinguished between the transport on surfaces 310–330 K and that on 340–360 K. On the lower surfaces he observed significant isentropic exchange between stratosphere and troposphere throughout all seasons. On the upper surfaces he observed a seasonal cycle in the transport with little exchange in the winter hemisphere. In section 3 we discuss in detail the seasonal variation in strength and position of the tropopause transport barrier revealed by the effective diffusivity field.

In section 4 we calculate effective diffusivity directly from the potential vorticity (PV) fields, rather than a “test tracer.” We then compare the position of the tropopause barrier defined by the minimum effective diffusivity with that determined more conventionally using PV values.

The lower limit of the vortex-edge barrier discussed in part 1 sometimes enters the region under consideration here in part 2. McIntyre [1995] called the region below the lower limit of the vortex-edge barrier the “sub-vortex.” He estimated that the sub-vortex transition occurred at about 400 K and noted evidence for less restricted mixing on isentropic surfaces below this level. This evidence is provided both by chemical data [Tuck, 1989; Podolske et al., 1989] and by contour advection studies [Chen, 1994]. In contrast, in part 1 no vortex-edge barrier was observed below 550 K in the Northern Hemisphere during some winter/spring months, implying a much higher sub-vortex transition. We consider this issue further in section 5.

Finally, we summarize and discuss the results presented here, and in part 1, in section 6 and present an objectively determined successor to the subjectively determined schematic Figure 1 of part 1, giving a more complete description of the isentropic transport and mixing structure of the troposphere and stratosphere between 300 and 850 K.

2. Effective Diffusivity

In part 1, we described a method to use “effective diffusivity” [Nakamura, 1996; Winters and Y’Asami, 1996] as a diagnostic of the transport and mixing structure of atmospheric flows.

The method is based on numerical solution of the two-dimensional advection-diffusion equation for a passive “test tracer” on a spherical surface, where the advection is provided by isentropic wind fields. We use a single-layer spectral model driven by 6-hourly European Centre for Medium-Range Weather Forecasting (ECMWF) initialized operational data from the period November 1996 to May 1998. The tracer field is output from the simulation at daily intervals, and for each output, the effective diffusivity $\kappa_{\text{eff}}$ is calculated as a function of equivalent latitude $\phi_e$, following the method described in part 1 section 2.2.

In part 1, section 2.3 and section 2.4, we presented results from a number of simulations designed to examine the issue of whether $\kappa_{\text{eff}}$ can be regarded as a diagnostic of the transport and mixing properties of the flow (rather than a particular tracer field). Given the substantial differences in flow structure between the upper-stratosphere/lower-stratosphere region to be examined here, and the stratospheric region considered in part 1, we re-examine this issue using winds for the 350 K surface rather than for the 450 K surface.

First, we examine again whether or not the tracer adjusts to a configuration where $\kappa_{\text{eff}}$ is independent of the initial conditions. As in part 1, two simulations were carried out, both initialized with a tracer profile equal to $\sin(\phi)$, the first beginning November 1, 1997 and the second beginning December 1, 1997, and both continuing until February 28, 1998. Figure 1a shows $\kappa_{\text{eff}}(\phi_e)$ calculated from the tracer fields from the two simulations on December 6, 1997, January 1, 1998, and February 28, 1998. A 10-day running-mean was applied to the $\kappa_{\text{eff}}$ (as in part 1, and as for the time evolution of $\kappa_{\text{eff}}$ to be shown later in Plate 3), and then the percentage difference (averaged over all $\phi_e$) in $\kappa_{\text{eff}}$ between the two simulations was calculated. The time evolution of this percentage difference is shown in Figure 1b. As at 450 K, the $\kappa_{\text{eff}}$ distributions for the two simulations are already similar on December 6, 1997, and show very close agreement at the later times, January 1, 1998 and February 28, 1998. The time evolution shown in Figure 1b confirms that, as at 450 K, after a “spin-up” time of 1 month there is always less than 5% difference between the two simulations.

The simulation that was begun November 1, 1997, was repeated with the different initial conditions for the tracer field used in part 1 (see Figure 3a of part 1 for graphs of these initial tracer conditions), chosen to give high and low gradients in different regions. Figure 2 shows that on December 1 (after the 1-month spin-up) the corresponding $\kappa_{\text{eff}}$ differ only in small-scale features.

On the basis of these results we argue that, as for the stratospheric levels considered in part 1, after a spin-up time of 1 month the effective diffusivity is independent of the tracer initial condition and may be regarded as a diagnostic of the properties of the flow.

In part 1 the results from selected simulations with different values of $\kappa$ were used to conclude that for the stratosphere the structure of $\kappa_{\text{eff}}$ is largely independent of $\kappa$, at least for $\kappa$ within a certain numerical range. Furthermore
The dependence of the actual values of $\kappa_{\text{eff}}$ on $\kappa$ appeared to be consistent with heuristic arguments.

We repeated these simulations using wind fields for the 350 K isentropic surface for the period December 1, 1997 to February 28, 1998. The values of diffusivity $\kappa$ were taken, respectively, to be $1.62 \times 10^5$ m$^2$ s$^{-1}$, $3.24 \times 10^5$ m$^2$ s$^{-1}$, and $4.8 \times 10^5$ m$^2$ s$^{-1}$, corresponding to damping times for the smallest resolved scale of 0.4, 0.5, and 0.1 days, respectively. Figure 3 shows the variation of $\kappa_{\text{eff}}$ with $\phi_e$ in the three simulations for one particular day, February 9, 1998. The latitudinal pattern of high and low values of $\kappa_{\text{eff}}$ is very similar between the three distributions and we therefore conclude that, within this range, the exact choice of $\kappa$ is relatively unimportant.

The simulations reported in part 1 used winds at T42 resolution, which have been argued to be more than sufficient for particle, contour, or tracer advection in stratospheric flows [e.g., Waugh and Plumb, 1994]. Near the tropopause, smaller-scale features in the winds are potentially more important than in the stratosphere, and consequently, it may be
the case that T42 resolution is not sufficient. To examine the sensitivity of the results to the truncation of the ECMWF wind fields, two simulations were performed for the period May 1, 1997 to August 31, 1997, the first using T106 winds and the second using T42 winds, and both using T106 resolution for the tracer field. Examples of the corresponding \( \kappa_{\text{eff}}(\phi_e) \) distribution are given in Figure 4a for June 11, 1997 and in Figure 4b for August 25, 1997, and are extraordinarily similar.

For the remainder of the paper we report results from simulations at T85 resolution with \( \kappa = 3.24 \times 10^5 \) m\(^2\) s\(^{-1}\), using winds at T42 resolution. A set of 4-month integrations were

**Figure 2.** The \( \kappa_{\text{eff}}(\phi_e) \) distribution on December 1, 1997, from five simulations using winds from 350 K, beginning November 1, 1997, with the tracer initial conditions shown in Figure 3a of part 1.

**Figure 3.** The \( \kappa_{\text{eff}}(\phi_e) \) distribution on February 9, 1998, for three simulations with different values of \( \kappa (1.62 \times 10^5 \) m\(^2\) s\(^{-1}\) dots, 3.24 \times 10^5 \) m\(^2\) s\(^{-1}\) solid, and 6.48 \times 10^5 \) m\(^2\) s\(^{-1}\) dashes), each using winds on the 350 K surface and using an initial condition of \sin(\phi) for the tracer on December 1, 1997.
Plate 1. The $\tilde{K}_{\text{eff}}(\phi, \theta)$ distribution, averaged over each 2 months during the period December 1996 to November 1997, and superimposed contours of zonal-mean zonal wind (as a function of latitude $\phi$, and $\theta$), with easterlies dashed and westerlies solid.
Plate 2. Graphs of $\tilde{\kappa}_{\text{eff}}(\phi_n)$ for various isentropic surfaces (different colors) in the range 300–390 K. Seasonal averages and the annual average are plotted.
Figure 4. The $\kappa_{\text{eff}}(\phi_e)$ distribution on (a) June 11, 1997 and (b) August 25, 1997 for two simulations beginning May 1, 1997 using isentropic winds from 350 K, one at T42 and the other at T106 horizontal resolution, and both using T106 horizontal resolution for tracer.

performed for each isentropic surface with the following potential temperatures $\theta$: 300, 310, 320, 330, 340, 350, 360, 370, 380, 390, 400, and 450 K. For each integration an initial spin-up time of 1 month was allowed before $\kappa_{\text{eff}}$ was calculated. As in part 1, the results will be presented in the form of the nondimensionalized quantity $\bar{\kappa}_{\text{eff}} - \kappa_{\text{eff}} / \kappa$, which takes the value 1 when a tracer contour is geometrically equivalent to a latitude circle.

Plate 1 shows the effective diffusivity for all equivalent latitudes on isentropic surfaces 300–450 K in the upper troposphere and lower stratosphere for the period December 1996 to November 1997. The zonal mean zonal winds are superimposed, plotted against latitude rather than equivalent latitude, to give a rough indication of the associated flow structure. Each panel shows $\kappa_{\text{eff}}$ averaged over 2 month periods and the six panels cover an entire annual cycle. The spatial variations in $\kappa_{\text{eff}}$ are less extreme in this region than higher in the stratosphere (part 1, Plates 1, 2 and 3) with much of the region being represented by moderate to high values of $\kappa_{\text{eff}}$ ($\ln \bar{\kappa}_{\text{eff}} > 1.5$) indicative of significant mixing.
3. Tropopause

All panels of Plate 1 show local minima in effective diffusivity within the range 370–400 K, indicating "subtropical-jet barriers." Parts of these barriers lie on "middleworld" isentropic surfaces (370–380 K) which connect the stratosphere and the troposphere, and are clearly associated with the tropopause. We will identify the "tropopause barrier" with the minimum value of $\kappa_{eff}$ on each isentropic surface below 380 K. We now consider the vertical variation and seasonal evolution of $\kappa_{eff}$, shown in Plate 1, below 400 K.

In winter (December–January in the Northern Hemisphere, June–July in the Southern Hemisphere) there are low values ($0.5 < \ln \kappa_{eff} < 1.5$) in both hemispheres, indicating a significant barrier, on isentropic surfaces in the range 330–390 K and equivalent latitudes 20°–40°. The lowest values of $\kappa_{eff}$ associated with this barrier are at the core of the subtropical jet at 350 K and at $\phi_e = 30^\circ$. In the Southern Hemisphere this barrier tends poleward with height to merge with the vortex-edge barrier at about 400 K. The tropopause barrier weakens and tends poleward at lower values, to be centered on about 40° at 320 K. Values of $\kappa_{eff}$ associated with the barrier are higher in the Northern Hemisphere winter than in the Southern, indicating that the barrier is weaker there.

As discussed in part 1, the transport and mixing structure of the winter stratosphere comprises a vortex and a surf zone. It was observed in part 1, and will be discussed further here in section 5, that the low values of $\kappa_{eff}$ associated with the vortex-edge barrier descend to a lower level in the Antarctic than in the Arctic: indeed, low values ($\ln \kappa_{eff} < 1$) of $\kappa_{eff}$ can be observed in the Southern Hemisphere high latitudes down to 380 K in June–July here in Plate 1. At lower levels, on the poleward side of the subtropical jet, moderately high $\kappa_{eff}$ values ($\ln \kappa_{eff} > 1.5$) indicate that mixing occurs up to a height determined by the position of the bottom of the vortex, i.e., in the Southern Hemisphere up to about 370 K in June–July and in the Northern Hemisphere up to about 420 K in December–January. On the equatorward side of the jet there are relatively high $\kappa_{eff}$ values ($1.5 < \ln \kappa_{eff} < 2$ in the south, $2 < \ln \kappa_{eff} < 2.5$ in the north) indicating mixing in a narrow subtropical band extending from about 360 K up to the stratospheric surf zone.

As the winter progresses into spring (February–March in the Northern Hemisphere, August–September in the Southern Hemisphere), the polar vortex descends farther. As a result, mixing is inhibited in the extratropics, above 380 K in the Northern Hemisphere and above 330 K in the Southern Hemisphere. In spring (April–May in the Northern Hemisphere, October–November in the Southern Hemisphere) the retreat of the polar vortex barrier allows mixing throughout the Northern extratropics. In the Southern Hemisphere, on the other hand, the polar vortex persists, inhibiting mixing above 360 K.

In summer (June–September in the Northern Hemisphere, December–March in the Southern Hemisphere) the values of $\kappa_{eff}$ are almost everywhere significantly higher than in winter. This is perhaps surprising for the tropopause since the upward Eliassen-Palm (EP) flux associated with the baroclinic eddies in the summer hemisphere is less than in the winter [Randel, 1992]. One explanation may be that the eddies have a greater mixing ability in the weaker zonal winds of the summer hemisphere than in the winter hemisphere. There is particularly strong mixing ($\ln \kappa_{eff} > 2$) at upper levels (above 350K) on the tropical side of the jet (especially June–September in the Northern Hemisphere and February–March in the Southern Hemisphere). This mixing is thought to be associated with the monsoon circulations. The tropopause barrier is seen to be substantially weakened in the summer, particularly in the Northern Hemisphere, probably as a result of the monsoon-related mixing [Chen, 1995; Dunkerton, 1995; Dethof et al., 1999]. This is consistent with the results, mentioned in section 1, of Chen [1995]. The interhemispheric difference is also consistent with observed interhemispheric differences in lower stratospheric water vapor [Pan et al., 1997; Rosenlof et al., 1997; Dethof et al., 1999]. Above about 360 K there appears to be no identifiable tropopause barrier in summer in either hemisphere.

In autumn (October–November in Northern Hemisphere and April–May in Southern Hemisphere) there is a general reduction in the values of $\kappa_{eff}$, indicating weaker mixing and an increase in strength of the tropopause barriers. However, in both hemispheres are weaker than in spring. For example, the northern tropopause barrier in October–November is weaker ($\ln \kappa_{eff} > 1$) than that in April–May ($\ln \kappa_{eff} < 1$) (even though the zonal winds show that the jet is stronger in October–November).

The low-level behavior is difficult to discern from Plate 1, but is clearer in the graphs of $\kappa_{eff}(\phi_e)$ shown in Plate 2. These graphs show that in most seasons there is a significant midlatitude minimum in $\kappa_{eff}$ down to and including the 320 K level; however, at 310 K the minimum is extremely weak and at 300 K there is no minimum at all. This indicates that, in both hemispheres, there is essentially no tropopause barrier below 320 K. The disappearance of the barrier is presumably associated with synoptic-scale baroclinic eddies whose mixing effect is expected to be strongest near the surface [e.g., Pierrehumbert, 1995].

The seasonal evolution is particularly clear from Plate 3, which shows $\kappa_{eff}(\phi_e,t)$ on the 310, 330, 350, and 370 K levels. For example, it can be seen that at 310 K the mixing is strongest in the Northern Hemisphere summer, consistent with evidence noted by Chen [1995]. At 330 and 350 K the tropopause barrier is visible at about $\phi_e = 35^\circ$ in winter, weakening and shifting poleward to about $\phi_e = 50^\circ$ in early summer. This summertime poleward displacement is consistent with the summertime poleward displacement of the subtropical jet (in the Northern Hemisphere the barrier is so weak in summer that it is difficult to identify its precise location.) At 370 K a strong tropopause barrier is observed at about $\phi_e = 30^\circ$ during winter, but it almost disappears during summer in both hemispheres. A distinct polar vortex barrier is visible during late winter at 350 and 370 K in the Southern Hemisphere (see section 5 for further discussion).
Interannual differences in the strength of the tropopause barrier can be seen, with a stronger barrier in the Northern Hemisphere winter 1997/1998 than in 1996/1997. Interannual differences can also be seen in the summer hemisphere subtropics, with stronger monsoon-related mixing, at 350 and 370 K in the Southern Hemisphere, in summer 1996/1997 than in summer 1997/1998. There is a corresponding difference in the tropopause barrier, which is weaker and further poleward (50° as compared to 40° at 350 K) in 1996/1997 than 1997/1998. There may be an interesting link here to interannual variability of the tropical tropospheric circulation, for example, that associated with El Niño.

As noted earlier, $\kappa_{\text{eff}}$ is a measure of part of stratopause-troposphere exchange, in particular the “mass-exchanging” part as opposed to the “mass-carrying” part that is measure by net mass fluxes [Holton et al., 1995]. The seasonal variation in the distribution of effective diffusivity therefore suggests a corresponding seasonal variation in stratopause-troposphere exchange. For example, the higher values of $\kappa_{\text{eff}}$ associated with the tropopause barrier in summer suggest more exchange. Appenzeller et al. [1996b] have previously noted the seasonal variation of the net mass flux across the extratropical tropopause, defined by the 2 PVU surface. They showed that the net mass flux peaked in late spring, consistent with earlier observations of radioactivity resulting from atmospheric testing of nuclear weapons. It is interesting that $\kappa_{\text{eff}}$ shows the mass-exchanging isentropic flux to be large in late spring, particularly in the Northern Hemisphere, and this may therefore also explain the radioactivity observations.

Bithell and Gray [1997] used contour advection to investigate the barrier effect of the tropopause in the Northern Hemisphere during October 1990. Consistent with our $\kappa_{\text{eff}}$ results, they show that at 310 K parcels of tropospheric origin, with PV less than 1 PVU, are able to mix into the stratosphere and all the way to the pole over 6 days. At 340 K they found a resistance to mixing of parcels with PV less than 1 PVU (of tropospheric origin) with parcels with PV greater than 4 PVU (of stratospheric origin). However, when they investigated contour stretching rates on the 310, 320, 330, and 360 K isentropic surfaces, they were unable to find a contour with minimum stretching rate that they could associate with the tropopause. The intrinsic limitations of the contour-based approach have been noted in section 1. Nevertheless, Plate 3 shows that Bithell and Gray might have obtained clearer results if they had chosen a different time to calculate their stretching rates, since in October the tropopause barrier in the Northern Hemisphere is weaker (as compared to January, for example).

Plate 1 shows a strong tropical barrier below 340 K and above 380 K (sometimes reaching as low as 360 K). The upper barrier was seen in the tracer calculations of Chen [1995], who noted that at 320 K it appeared to be stronger than the tropopause barrier. The very substantial difference in the strengths of the two barriers at the lower levels can be seen clearly here. The lower tropical barrier, indicated by the triangular region of low $\kappa_{\text{eff}}$ with base 20°N–20°S at 300 K and top at 340 K, seems to be associated with strong zonal flow. (Inspection of the streamfunction field shows that zonal asymmetries are weak.) The higher values of $\kappa_{\text{eff}}$ seen in the tropical troposphere at 350 K are likely to be caused by a combination of slightly weaker zonal flow and strong anticyclones. The upper tropical barrier, indicated by a second triangular region of low $\kappa_{\text{eff}}$, seems to be associated with the more zonally symmetric flow above 360 K and joins to the lower stratospheric tropical reservoir discussed in section 3.5 of part 1.

The seasonal evolution of the tropical barrier over the period December 1996 to May 1998 may be seen in Plate 3. Note that both lower (330 K) and upper (370 K) barriers were much weaker during the first half of 1997 than at other times. Again, there is an interesting possible link to interannual variability in the tropical tropospheric circulation.

4. Effective Diffusivity Calculated From Potential Vorticity

4.1. Troposphere and Lower Stratosphere

Effective diffusivity may be calculated from PV derived from observations, just as it can from model tracer or observed chemical species. In the extent that PV is a conserved tracer, we might expect the same features to be shown by effective diffusivities calculated from PV and from model or chemical tracers. We now compare effective diffusivity ($\kappa_{\text{eff}}^{PV}$) calculated from PV fields derived directly from ECMWF data at 142 resolution with the effective diffusivity ($\kappa_{\text{eff}}^{PV}$) calculated from our “test tracer” and discussed earlier.

Plate 4 shows the time-latitude variation of $\kappa_{\text{eff}}^{PV}$ on the 330, 350, and 370 K levels, and may be compared with Plate 3. Note that the blobs of anomalously high $\kappa_{\text{eff}}^{PV}$ seen in this plate, particularly at lower levels, apparently arise from inaccuracies in the data analysis, and we therefore do not consider them to be significant. (We also calculated $\kappa_{\text{eff}}^{PV}$ on the 310 K level, but the anomalously high values of $\kappa_{\text{eff}}^{PV}$ occurred sufficiently frequently that we do not show the results here.) Ignoring the blobs of high $\kappa_{\text{eff}}^{PV}$, there are strong similarities between the structures seen in this plate and those seen in Plate 3. In particular, as in Plate 3, a tropopause barrier, with low values of $\kappa_{\text{eff}}^{PV}$, is visible at all levels in the subtropics of the winter hemisphere, and these barriers weaken (i.e., values of $\kappa_{\text{eff}}^{PV}$ increase) and shift poleward in early summer. The tropopause barriers are seen to be weaker at 330 K than at higher levels. Also, a distinct polar vortex barrier is visible during late winter at 350 and 370 K in the Southern Hemisphere.

The differences that exist between the absolute values of $\kappa_{\text{eff}}^{PV}$ shown in Plate 4 and of $\kappa_{\text{eff}}$ shown in Plate 3 are not very significant since they depend on the spatial scales resolved in the two fields. (The resolution of the PV field is nominally T42 but may well be coarser than that in practice: the tracer simulation was at T85 resolution.) However, there are important qualitative differences between Plates 4 and 3, most substantially in the tropics. For example, on 330 K
the broad region of low $\kappa_{\text{eff}}$, identified in section 3 as the lower tropical barrier, is absent in the $\kappa_{\text{eff}}$ distribution. Another region of noticeable difference is in the subtropics of the summer hemisphere at 450 and 310 K, where the high values of $\kappa_{\text{eff}}$ seen in Plate 3, thought to indicate monsoon-related mixing, are generally absent in Plate 4. These differences may be due to poor-quality PV data, or to the fact that diabatic processes affect the PV field strongly in the tropics and subtropics (whereas they do not directly affect tracer advected on isentropic surfaces). There also appear to be greater interhemispheric differences, with Plate 4 showing systematically higher values of $\kappa_{\text{eff}}$ at high latitudes in the Northern Hemisphere than in the Southern Hemisphere. This may be a consequence of poor-quality PV data in the ill-observed Southern Hemisphere.

4.2. Tropopause Definition

It is now conventional to define the tropopause by a particular value of PV [e.g., Holton et al., 1995]. A wide range of PV values have been used as this definition. The values proposed include 1 PVU [Shapiro, 1978], 2 PVU [Appenzeller et al., 1996a], 3 PVU [Speth et al., 1994], and 3.5 PVU [Hosler et al., 1991, 1993], and it is unclear which of these, if any, is optimal.

The results that we have presented above give the possibility of an alternative definition of the tropopause based on the location of the relevant local minimum in effective diffusivity (either $\kappa_{\text{eff}}$ or $\kappa_{\text{eff}}^\text{PV}$), i.e., the location of the “tropopause barrier”, on each isentropic surface.

We implemented this definition as follows. We calculated time series of $\kappa_{\text{eff}}^\text{PV}$ at 6 hourly intervals on isentropic surfaces 330, 350 and 370 K, we then identified the minimum $\kappa_{\text{eff}}^\text{PV}$ on each surface, within a suitable range of equivalent latitudes (to avoid other minima associated either with the vortex-edge or the tropical barrier). The calculation was then repeated using $\kappa_{\text{eff}}$ (but calculated at daily rather than 6-hourly intervals).

Figure 5 shows graphs of equivalent latitude ($\Phi_e^\text{PV}$) at the minima in $\kappa_{\text{eff}}^\text{PV}$, as functions of time, on the 330, 350, and 370 K isentropic levels. Superimposed are the corresponding graphs of equivalent latitude ($\Phi_e$) at the minima in $\kappa_{\text{eff}}$. Many features in the two graphs are similar. Both show, at all levels, in 1997 the poleward migration of the tropopause in Southern Hemisphere late summer and in Northern Hemisphere late summer. Both also show that there is much weaker poleward migration of the tropopause in Southern Hemisphere late summer 1998. The most substantial differences between the graphs appear on the 330 K surface, where the minima in effective diffusivity are weaker and therefore discrimination between the tropopause minimum and other higher- or lower-latitude minima is less straightforward. Difficulty in identifying the position of the minimum effective diffusivity when the barrier is very weak is also the likely explanation for the differences between $\Phi_e^\text{PV}$ and $\Phi_e$ in the Southern Hemisphere in autumn.

To compare this new definition of the tropopause with one based on a particular value of PV, we show in Figure 6 graphs of the value of PV corresponding to the minimum $\kappa_{\text{eff}}^\text{PV}$, calculated from the 6-hourly data, with the monthly running-mean superimposed. It can be seen that the values fall predominantly in the range 1–6 PVU; however, the value is seen to fluctuate dramatically within this range. Therefore it seems impossible to select a single value of PV that corresponds to the “tropopause barrier” definition at all levels. There seems, instead, to be closest correspondence to $\pm 2$ PVU at 330 K, and perhaps $\pm 2.5$ PVU at 350 K and $\pm 4.5$ PVU at 370 K. Deviations from these values, which are sometimes quite substantial, seem predominantly associated with the monsoons (January–March in the Southern Hemisphere, July–September in Northern Hemisphere) when typically the PV value is higher.

5. Sub-vortex

As pointed out above in section 3, the seasonal evolution of the lowermost part of the polar vortices may also be followed in Plate 1. Looking at the Southern Hemisphere, in June–July at high latitudes above 380 K a region of very low $\kappa_{\text{eff}}$ (in $\kappa_{\text{eff}} < 0.5$), first observable in April–May, is seen to be closely associated with the core of the polar-night jet. This region of very low $\kappa_{\text{eff}}$, which indicates a significant barrier to transport between the polar region and midlatitudes, is the lower extent of the Antarctic vortex-edge barrier discussed in part 1. During August–November the vortex and the associated region of very low $\kappa_{\text{eff}}$ both descend.
Plate 3. The $\kappa_{\text{eff}}(\phi, t)$ distribution for December 1996 to May 1998 on the 310, 330, 350, and 370 K isentropic surfaces. Superimposed are contours of zonal-mean zonal winds (as a function of $\phi$ and $t$), with easterlies dashed and westerlies solid. A 10-day running-mean has been applied to the daily $\kappa_{\text{eff}}$ and wind data.
below 380 K. The persistence of the vortex-edge barrier into November at levels 400–500 K was noted in section 3.1 of part 1 (see Plates 2 and 3). If we define the vortex-edge barrier to be the region of low $\kappa_{\text{eff}}$, say $\ln \kappa_{\text{eff}} < 1$, then the $\kappa_{\text{eff}}$ distribution suggests that the sub-vortex transition, marked by the lower limit of this barrier, occurs at about 380 K in June–July and October November. In August–September the sub-vortex transition occurs as low as 350 K. This is consistent with the contour-advection results of Chen [1994], who argues that there is a strong vortex-edge barrier in August 1993 down to 375 K. McIntyre [1995] notes that the wind fields used by Chen may not include a good representation of relatively small-scale eddies near the Antarctic continent and therefore may underestimate the mixing at low levels, and the same criticism might well apply to our results. However, we use ECMWF winds which have higher horizontal and vertical resolution than the U.K. Meteorological Office (UKMO) winds used by Chen and which therefore should give improved results in the lower-stratosphere/upper-troposphere region relevant here.

The situation in the Northern Hemisphere is rather different. The bottom of the Arctic vortex-edge barrier is also visible in Plate 1; however, it descends less, and for a shorter period, than the Antarctic vortex-edge barrier. In particular, during February–March the very low values of $\kappa_{\text{eff}}$ ($\ln \kappa_{\text{eff}} < 0.5$) descend only to about 400 K, and in fact, during December–January no very low values of $\kappa_{\text{eff}}$ are observed at all below 450 K. The $\kappa_{\text{eff}}$ distribution suggests that, unlike the Southern Hemisphere winter/spring mentioned above, there is no barrier to transport from polar regions to midlatitudes at levels between 450 and 380 K during some of the winter/spring months December–May, and that the sub-vortex transition generally occurs at a considerably higher level in the Northern Hemisphere than in the Southern Hemisphere. There is, of course, strong interannual variability in the Northern Hemisphere and inspection of Plate 1 of part 1 suggests that throughout most of the following Northern Hemisphere winter/spring (1997/1998) the sub-vortex transition occurred at a high level, with the vortex-edge barrier descending below 450 K only in February.

Therefore, for neither hemisphere does the level of 400 K, previously identified as the transition from vortex to subvortex, seem particularly significant. Instead, the transition appears to vary considerably throughout the season between about 350 and 380 K in the Southern Hemisphere, and between about 400 K and above 450 K in the Northern Hemisphere.

Throughout the period June to November the Antarctic vortex-edge barrier joins via a region of relatively low $\kappa_{\text{eff}}$ ($1 < \ln \kappa_{\text{eff}} < 1.5$) to a minimum in $\kappa_{\text{eff}}$ ($\ln \kappa_{\text{eff}} < 1$) associated with the subtropical jet. This minimum has been referred to earlier as the subtropical-jet barrier, below 380 K it corresponds to the tropopause barrier.

The transport and mixing properties of the flow deduced above from the $\kappa_{\text{eff}}$ distribution are consistent with the contour lengthening calculations, using winds from the UKMO dataset, for the Antarctic sub-vortex of Chen et al. [1994]. They considered the 375 K surface during August 1992; however, the lack of substantial interannual variability in the
Plate 4. The $\tilde{\kappa}_\text{eff}^{PV}(\phi_e, \theta)$ distribution for December 1996 to May 1998 on the 330, 350, and 370 K isentropic surfaces. The $\kappa^{PV}_\text{eff}$ distribution was calculated from PV fields derived from ECMWF data at 6-hourly intervals. Superimposed are contours of zonal-mean zonal winds (as a function of $\phi$ and $t$), with easterlies dashed and westerlies solid. A 10-day running-mean has been applied to the daily $\kappa^{PV}_\text{eff}$ and wind data.
Plate 5. The seasonal and annual averages of the effective diffusivity distribution $\kappa_{\text{eff}}$, originally calculated as a function of equivalent latitude and potential temperature, re-displayed as a function of equivalent latitude and log-pressure height. (Zonal-mean temperature data have been used to assign the potential temperatures (300–850 K) at each equivalent latitude to log pressure heights.)
Antarctic vortex gives some confidence in the usefulness of comparing their results with those above. Chen et al. initialized contours in the region 70°S to 30°S and found that those initialized at 70°S (which would, according to Plate 1, be just inside the vortex) and at 30°S (according to Plate 1 just equatorward of the core of the subtropical jet) were stretched considerably more, over a period of 20 days, than intermediate contours. Chen et al. also showed that neither of these two extreme contours crossed 50°S. These results suggest a broad barrier to transport. However, the picture shown here for August–September clearly demonstrates that it is an oversimplification to consider this broad barrier to be entirely associated with the vortex edge. The subtropical jet has an important influence too, and it is the overlap of the vortex-edge and subtropical-jet barriers which results in a broad region of low mixing.

In the tropics the 400 K isentrope is close to the absolute upper limit of the tropical convection and hence is one possible definition of the tropical tropopause [e.g., Holton et al., 1995]. It has sometimes been speculated that there is some significance in the fact that the tropical tropopause and the vortex to sub-vortex transition at high latitudes lie on the same isentropic surface. Our results, which show the vortex to sub-vortex transition occurring sometimes above 450 K and other times as low as 350 K suggest that there is unlikely to be any close connection between these two features.

6. Discussion

In this paper we have examined the effective diffusivity field, calculated from an advected tracer, on isentropic surfaces in the range 300–450 K. This has allowed us to identify a tropopause transport barrier and to characterize the transport structure of the sub-vortex region. In particular, we have noted that the tropopause barrier is much stronger in the winter hemisphere than in the summer, when it shifts poleward and is substantially weakened by monsoon mixing. We also noted that the sub vortex transition is on average substantially higher (around 120 K, or 20 km) in the Northern Hemisphere than in the Southern Hemisphere (around 370 K, below 15 km).

The minimum value of effective diffusivity in each hemisphere on certain isentropic surfaces provides a natural definition of the extratropical tropopause, consistent with the idea of the extratropical tropopause as an eddy-transport barrier [e.g., Holton et al., 1995]. This is arguably a more fundamental and general definition than one based on a particular value of PV or a particular value of lapse rate. The distinction between the different definitions may be particularly important in considering future long-term changes in tropopause position, or in general circulation model studies such as those reported by Thuburn and Craig [1997] where the changes in tropopause position due to drastic changes in parameters such as rotation rate or surface temperature are being investigated.

The results presented here, together with those in part 1, have given a complete picture of the height latitude variation and seasonal evolution of the transport barriers and mixing regions identified previously in the schematic Figure 1 in part 1.

To allow comparison with the schematic figure, the $\nu_{\text{eff}}$ fields shown here and in part 1 are displayed as a function of equivalent latitude and log pressure height in Plate 5. (Zonal mean temperature data have been used to associate the potential temperatures (300–850 K) at each equivalent latitude to a log-pressure height.)

The picture for the annual average shows the following: in the extratropics in the troposphere, below 350 K (17 km), a region of strong mixing; in the midlatitudes at about 40° centered on the 550 K surface, a barrier associated with the tropopause; at high latitudes in the stratosphere, in each hemisphere, a region of weak mixing associated with the winter polar vortex; and in the tropics, in the stratosphere, a broad region of weak mixing associated with the tropical reservoir. The tropical reservoir appears broadest and the mixing is weakest at about 550 K (23 km), as is consistent with previous studies mentioned in part 1 section 3.5.

Details of the seasonal variation of the transport have been discussed in the relevant sections of parts 1 and 2, but in the following we re-emphasize points that are apparent from Plate 5. The winter (DJF for Northern Hemisphere, JJA for Southern Hemisphere) shows strong tropopause and polar vortex barriers. In the Northern Hemisphere the vortex barrier extends to 60° and is seen to be weak between 450 K (20 km) and 600 K (25 km) and strong above this. In the Southern Hemisphere the barrier extends further equatorward, to 50°, and is strong above 400 K (16 km). Also, the lower values of $\nu_{\text{eff}}$ indicate the barrier is stronger than in the Northern Hemisphere. Mixing is observed within the vortex above 600 K (25 km) in the Southern Hemisphere. In spring (MAM for Northern Hemisphere, SON for Southern Hemisphere) there is little sign of a vortex barrier in the Northern Hemisphere, but the greater persistence of the vortex in the Southern Hemisphere means that a significant barrier may be seen there, especially at lower levels (400–600 K, 16–35 km). The tropopause barrier remains strong in both hemispheres during this season. In summer (JJA for Northern Hemisphere, DJF for Southern Hemisphere) the tropopause barrier is seen to be very weak in both hemispheres and there is strong mixing in the subtropics between 350 and 400 K (12–16 km). The summer extratropical mixing extends up into the stratosphere, to 600 K (25 km) in the Northern Hemisphere and higher in the Southern Hemisphere. In autumn (SON for Northern Hemisphere, MAM for Southern Hemisphere) the tropopause barrier is seen to be stronger than during the summer, although still weaker than in winter. In the stratosphere a vortex barrier can be observed above about 600 K (25 km) in the Northern Hemisphere and above 450 K (20 km) in the Southern Hemisphere. Finally, comparing the DJF and JJA pictures shows clearly the displacement of the tropical reservoir toward the summer pole.

In these two papers we have demonstrated that effective diffusivity is a powerful diagnostic for identifying transport barriers and mixing regions and comparing them in a quantitative manner. The challenge ahead is to relate the values
of effective diffusivity to actual fluxes of tracer. This would
open up the possibility of using effective diffusivity as a ba-
sis for a precise quantitative description of transport, both to
diagnose it from observations and to represent it in models.

Acknowledgments. We are grateful to David Tan, Warwick
Norton, Alan Iivi, Martin Chipperfield, Michael McIntyre, Noboru
Nakamura, and Iois Sreenman-Clark for advice and assistance with
this work. Financial support for this work was provided by the
U.K. Natural Environmental Research Council (in part by the U.K.
Universities Global Atmospheric Modelling Programme and in part
through a research studentship), the EC (through the TOASTE-C
project), and Trinity College, Cambridge.

References

Appenzeller, C., H. C. Davies, and W. A. Norton, Fragmentation
of stratospheric intrusions, J. Geophys. Res., 101, 1435–1456,
1996a.

Appenzeller, C., J. R. Holton, and K. H. Rosenlof, Seasonal vari-
ation of mass transport across the tropopause, J. Geophys. Res.,

Bithell, M. and J. I. Gray, Contour lengthening rates near the

Chen, P., Permeability of the Antarctic vortex edge, J. Geophys.

Chen, P., Isentropic cross-tropopause mass exchange in the exat-

Chen, P., J. R. Holton, A. O’Neill, and K. Swinbank, Quasi-
horizontal transport and mixing in the Antarctic stratosphere,

Dethof, A., A. O’Neill, J. M. Slingo, and H. G. J. Smit, A mecha-
nism for moistening the lower stratosphere involving the Asian
summer monsoon, Q. J Roy Meteorol. Soc., B 125, 1079–1106,
1999.

Dunkerton, T. J., Evidence of meridional motion in the summer
lower stratosphere adjacent to monsoon regions, J. Geophys.

Haynes, P. H., and E. F. Shuckburgh, Effective diffusivity as a di-
agnostic of atmospheric transport, 1. stratosphere, J. Geophys.
Res., this issue.

Hoerling, M. P., T. K. Schaack, and A. J. Lensen, Global objective

Hoerling, M. P., T. K. Schaack, and A. J. Lensen, A global analysis of
stratospheric-tropospheric exchange during northern winter,

Holton, J. R., P. H. Haynes, M. E. McIntyre, A. R. Douglass, R. B.
Rood, and I. Pfister, Stratosphere–troposphere exchange, Rev

McIntyre, M. E., The stratospheric polar vortex and sub vortex:
Fluid dynamics and midlatitude ozone loss, Phil. Trans. R. Soc.

Nakamura, N., Two dimensional mixing, edge formation, and per-
meability diagnosed in area coordinates, J. Atmos. Sci., 53,
1524–1537, 1996.

Pan I., S. Solomon, W. Randel, J. F. Lamarque, P. Hess, J. Gille,
E. W. Chio, and M. P. McCormick, Hemispheric asymmetries
and seasonal variations of the lowermost stratospheric water va-
por derived from SAGE II data, J. Geophys. Res., 107, 28177–
28184, 1997.

Pierrehumbert, R. T., Tracer microstructure in the large eddy dom-
inated regime, in Chaos Applied to Fluid Mixing, edited by
H. Arfe, and M. S. C. Naschie, pp. 347–365, Pergamon/Elsevier,

Pierrehumbert, R. T., and H. Yang, Global chaotic mixing on isen-

Podolske, J. R., Loewenstein, S. Strahan, and K. Chan, Strato-
spheric nitrous oxide distribution in the southern hemisphere,

Randel, W., Global Atmospheric Circulation Statistics, 1000–1mb.

Rosenlof, K. H., A. F. Tuck, K. K. Kelly, J. M. Russell III, and
M. P. McCormick, Hemispheric asymmetries in water vapor and
influences about transport in the lower stratosphere, J. Geophys.

Shapiro, M. A., Further evidence of the mesoscale and turbu-
ent structure of upper level jet stream-frontal zone systems,

Spaep, P., D. R. Johnson, and T. K. Schaack, Stratospheric-
tropospheric exchange during the Presidents’ Day storm, Mon.

Thuburn, J., and G. C. Craig, GCM tests of theories for the height

Tuck, A. F., Synoptic and chemical evolution of the Antarctic vortex
in late winter and early spring, 1987, J. Geophys. Res., 94,

Waugh, D. W., and R. A. Plumb, Contour advection with surgery:
A technique for investigating fine-scale structure in tracer trans-

Winters, K. B., and E. A. D’Asaro, Disccular flux and the rate of

P. H. Haynes and E. F. Shuckburgh, Centre for Atmospheric
Science, Dept Applied Mathematics and Theoretical Physics,
University of Cambridge, U.K. (P.H.Haynes@damtp.cam.ac.uk;
E.F.Shuckburgh@damtp.cam.ac.uk)

(Received August 20, 1999; revised January 26, 2000; accepted
February 4, 2000.)