Effective diffusivity as a diagnostic of atmospheric transport

1. Stratosphere

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Abstract. The transport and mixing properties of the isentropic flow in the lower and middle stratosphere are analyzed by using observed winds to advect a tracer on isentropic surfaces in the range 400--850 K. The effective diffusivity diagnostic introduced by Nakamura and collaborators is applied to the tracer field in order to identify barriers to transport and mixing regions, and to follow their seasonal evolution. Large effective diffusivity corresponds to strong mixing, and small effective diffusivity corresponds to weak mixing, i.e., to barriers. The effective diffusivity shows, in the winter stratosphere of each hemisphere, the evolution of the vortex-edge barrier and the midlatitude surf zone, and also the extent of any mixing within the vortex. At low latitudes in the stratosphere there is a region of low effective diffusivity whose latitudinal width varies with height, broadening substantially from 400 K to 550 K. The low values of effective diffusivity in this "tropical-reservoir" region imply little isentropic transport into or out of it. There is a strong seasonal cycle to the reservoir, which has different forms at 400 K, 450--600 K, and above 650 K, determined by the relative influences of tropospheric synoptic eddies and stratospheric planetary waves. Comparison of effective diffusivity between the Northern Hemisphere winters 1996/1997 and 1997/1998 shows strong differences at low latitudes according to the phase of the quasi-biennial oscillation (QBO). When there are QBO easterlies, there is a broad region of very low effective diffusivity at low latitudes. When there are QBO westerlies, there are very low values of effective diffusivity at low latitudes within the westerlies themselves but larger values at their edges.

1. Introduction

When considering transport of chemical tracers in the stratosphere, it has proved useful to divide the transport into components along isentropic surfaces ("isentropic transport") and across isentropic surfaces ("diabatic transport"). The distribution of long-lived chemical tracers is strongly influenced by both components. From the simplest viewpoint the diabatic transport, primarily associated with the large-scale Brewer-Dobson circulation, tends to steepen tracer contours in the height-latitude plane, which increases latitudinal gradients. The isentropic transport, primarily a dispersive transport associated with the eddies, has an opposite flattening effect, which reduces latitudinal gradients on isentropic surfaces [e.g., Andrews et al., 1987, chapter 9]. One might call this the "old paradigm" for stratospheric tracer transport. Under the old paradigm the natural way to quantify isentropic transport is by an eddy-diffusion coefficient, varying as a function of latitude, height, and time.

Over the last decade or so, new observations, from satellite instruments and from aircraft and balloons, have revealed strong spatial inhomogeneity in the distribution of stratospheric chemical tracers [e.g., McIntyre and Palmer, 1983, 1984; Batchart and Remsberg, 1986; Tick, 1989]. The most well known example of such inhomogeneity is the vortex/surf-zone structure in the winter stratosphere. The surf zone, within which there is rapid isentropic mixing, is bounded on its poleward side by a transport barrier at the vortex edge, across which there is only very slow isentropic transport. On its equatorward side, the surf zone is bounded by a subisentropic transport barrier. McIntyre [1993] calls these barriers "eddy-transport barriers," to emphasize the fact that they do not act as barriers to the zonally averaged Brewer-Dobson circulation.

Isentropic advection studies, in which particles or material contours are advected by large-scale winds taken from meteorological datasets, have given much insight into these spatially inhomogeneous transport regimes. Some studies have been associated with the "vortex-edge barrier" [Bowman, 1993; Chen, 1994; Chen et al., 1994b; Dahlberg and Bowman, 1994; Pierrehumbert, 1991; Plumb et al., 1994; Waugh et al., 1994], and others associated with the "subtropical barrier" [Waugh, 1996; Chen et al., 1994a].

These developments have motivated a "new paradigm" for stratospheric tracer transport. The stratosphere is partitioned into quasi-Lagrangian regions of limited latitudinal extent. Within each region, tracer concentrations are approximately
homogeneous on isentropic surfaces, because of rapid isentropic mixing by the eddies. Between these regions there are barriers to transport, across which there are sharp changes in tracer concentrations. In this new paradigm the natural way to quantify isentropic transport is by the permeability, varying as a function of height and time, of barriers, whose position varies in space and time. In reality, of course, both old and new paradigms are extreme simplifications, and the best approach to representing transport is probably to combine elements of both, as in the "tropical pipe" model of Plumb [1996].

Figure 1 [following Holton et al., 1995; World Meteorological Organization (WMO), 1999] shows schematically a possible partition of the troposphere and stratosphere into regions separated by eddy-transport barriers, indicated by thick curves. This partition provides a very useful framework for discussing transport, but it has been arrived at somewhat subjectively, based on a large set of different observations and numerical simulations, including those mentioned above. In this paper we describe a method of objectively identifying barriers to isentropic transport and of quantifying the permeability of such barriers.

One possible approach to identifying barriers is to consider particle separation, as motivated by the analogy with chaotic-advection flows [Pierrehumbert and Yang, 1993], in which regions of rapid mixing, characterized by exponentially fast separation of particles, are distinguished from barrier regions, where separation of particles is much slower. This has prompted a number of studies of atmospheric transport where separation rates of particles, or equivalently stretching rates of material contours, have been calculated by advecting particles or material contours in numerical simulations driven by observed winds or by winds from quasi-realistic models [Pierrehumbert, 1991; Pierrehumbert and Yang, 1993; Bowman, 1993; Norton, 1994; Waugh et al., 1994; Chen, 1994; Chen et al., 1994b]. This approach has been applied with some success to give an objective identification of the vortex edge barrier in the winter stratosphere in both the Northern Hemisphere [Waugh et al., 1994] and the Southern Hemisphere [Bowman, 1993; Chen, 1994].

The approach used in this paper is to diagnose the transport properties of atmospheric flows, defined by large-scale meteorological datasets, by solving the advection-diffusion equation for a "test tracer" in such flows, and then calculating the "effective diffusivity" diagnostic of Nakamura [1996] from the resulting tracer field. Effective diffusivity is a measure of the geometric structure of a tracer field. It is large where the geometric structure is complex, and
where the structure is simple. Mixing regions have
large stretching rates and hence produce complex geomet-
ic structure, implying large effective diffusivity. Barrier
regions have small stretching rates and hence produce sim-
ple geometric structure, implying small effective diffusiv-
ity. One possible application of effective diffusivity is that
adopted by Nakamura and Ma [1997], who use it to ana-
yze observed chemical tracer fields, in their case nitrous
oxide (N₂O) fields measured by the Cyrogenic 1.4m Array
Satalon Spectrometer (CLAES) instrument on the Upper
Atmosphere Research Satellite (UARS). Our approach is dif-
ferent in that we use the effective diffusivity calculated from
the “test tracer” to diagnose directly the isentropic transpor-
t and mixing properties of the velocity fields themselves.

The structure of the paper is as follows. Effective diffus-
vity and the details of methods used in this paper are de-
scribed in section 2. Then in section 3, the results of ap-
plying our approach to study the isentropic transport prop-
ties of the flow in the stratosphere are presented, first con-
sidering the extratropics and then the tropics. In section 4 we
present a summary and discussion of the results. In part 2
Haynes and Shuckburgh, this issue] we discuss the lower
stratosphere and upper troposphere. A further paper will
analyze interannual variability, in particular that associated
with the QBO.

The goal of this paper is to present a comprehensive pic-
ture of the height-latitude structure of the isentropic trans-
port and of its seasonal variation. (A first indication of possi-
ble interannual variability is also given.) The many previous
studies mentioned above have been of isolated subsections
of the troposphere and stratosphere, often considering par-
cular seasons of an individual year. Because a variety of
very different methods have been used, quantitative cross-
comparison of results between different height-latitude re-
gions and different seasons is difficult. The advantage of the
approach presented here is that the same technique is applied
on all high-latitude regions and seasons, thereby allowing
direct comparison and a quantitative assessment of the relative
strengths of the transport barriers.

2. Diagnostic Approach

2.1. Effective Diffusivity

Both Nakamura [1996] and Winters and D’Asaro [1996]
have introduced a quantity, “effective diffusivity,” for the
consideration of the evolution of a conservative chemical trac-
er with concentration c(x, t) according to a two-dimen-
sional advection-diffusion equation

$$\frac{\partial c}{\partial t} + u \cdot \nabla c = \nabla \cdot (\kappa \nabla c),$$  

(1)

where u(x, t) is the velocity field and \( \kappa \) is a constant diffu-
sivity. For each value \( \kappa \) of the concentration, the function
\( I(\kappa, t) \) may be defined as the area of the region for which
the tracer concentration c(x, t) is greater than or equal to \( \kappa \)
Butchart and Remsberg, 1986]. By definition, A(\kappa, t) is
a monotonic function of \( \kappa \), and there is therefore a unique
inverse function \( C(\kappa, t) \) such that \( C(A(\kappa, t), t) = \kappa \).

Nakamura and Ma [1997] showed that, providing the velocity
field u(x, t) is nondivergent, the function \( C(\phi, t) \) satisfies the
diffusion-only equation

$$\frac{\partial C(\phi, t)}{\partial t} = \frac{1}{r^2 \cos \phi_e} \frac{\partial}{\partial \phi_e} \left[ \kappa_{\text{eff}}(\phi, t) \cos \phi_e \frac{\partial C(\phi, t)}{\partial \phi_e} \right],$$  

(2)

where \( \kappa_{\text{eff}} \), the effective diffusivity, is defined by

$$\kappa_{\text{eff}}(\phi, t) = \kappa \frac{\langle |\nabla c|^2 \rangle}{\langle \nabla C/\partial \phi_e \rangle^2} = \frac{\kappa L_{\text{eq}}^2(\phi, t)}{2 \pi r \cos \phi_e}.$$  

(3)

Here \( \langle \cdot \rangle \) denotes the average over the area between ad-

cacent tracer contours and is defined by

$$\langle \cdot \rangle = \frac{\int \cdot \, dl}{\int \, dl},$$

where the integrals are around a contour of c. The quantity
\( L_{\text{eq}}(\phi, t) \), defined by the second equality in (3), is known
as the “equivalent length.” The transformation to \( \phi_e \) as an
independent variable removes the advective terms from the
evolution equation for the tracer concentration, and conse-

sequence (3) simply describes the diffusion of tracer relative to contours of \( \phi_e \). The equivalent length \( L_{\text{eq}}(\phi, t) \) can be related to the actual length \( L(\phi, t) \) of the tracer con-
tour corresponding to equivalent latitude \( \phi_e \) since, from the
Cauchy-Schwartz inequality,

$$L_{\text{eq}}^2 = L^2 \frac{1}{|\nabla c|} \geq L^2.$$  

(4)

Here \( \langle \cdot \rangle \) denotes the line average around a tracer contour and may be written as

$$\langle \cdot \rangle = \frac{\int \cdot \, dl}{\int \, dl},$$

where, again, the integrals are around a contour of c. Br-
dly speaking, the effective diffusivity is therefore largest
where the tracer contours are longest relative to their mini-
mum possible length \( 2\pi r \cos \phi_e \), i.e., where the geometric
structure of tracer contours is most complex. Since the mix-
ing ability of a flow is essentially the ability to produce com-
plex tracer contours, it follows that the effective diffusivity is
a measure of mixing ability. In the spatially inhomogeneous
flows of interest, the effective diffusivity will therefore tend
to be relatively large in mixing regions and relatively small
in barrier regions. This is illustrated by Nakamura [1996],
who shows the behavior of \( \kappa_{\text{eff}} \) calculated from the density
distribution in a simulation of nonlinear Kelvin-Helmholtz
instability. He finds, as expected, that \( \kappa_{\text{eff}} \) is largest in the
center of the shear layer, where there is strong mixing, and
is smallest at the edges of the shear layer, where mixing is
weak.
In a flow with time-varying mixing ability, the geometric structure of a tracer at any instant will depend on the history of the flow (weighted towards the recent past, defined by some “memory time”). The memory time will be shorter when \( \kappa \) is larger. Temporary changes in the mixing ability of the flow will be fully represented by \( \kappa_{\text{eff}} \) only if they persist for longer than the memory time.

The previously noted relation between particle separation rates and mixing ability allows effective diffusivity to be considered a hybrid Eulerian-Lagrangian diagnostic, which takes account of recent particle separation rates and assigns the values of the separation rates to particular quasi-geographical regions defined by values of the equivalent latitude \( \phi_e \).

2.2. Application to Atmospheric Flows

The focus in this paper is to use effective diffusivity to measure the time-varying transport and mixing properties of the large-scale isentropic flow. All of the results to be presented will be based on numerical solution of the two-dimensional advection-diffusion equation (1) on a spherical surface, where \( \mathbf{u}(\mathbf{x}, t) \) is the nondivergent part of the horizontal velocity field on a given isentropic surface, as specified by large-scale meteorological data. (Separate calculations including the divergent part of the velocities produce results with no discernible differences.)

The assumptions behind this general approach of using large-scale observed winds to advect particles or tracer have been discussed many times previously (see, for example, Waugh and Plumb [1994]). Particular details relevant here are as follows. The advection-diffusion equation was integrated using a single-layer spectral model. The diffusivity \( \kappa \) was always taken to be sufficiently large to avoid the problems sometimes associated with the use of the spectral technique. The velocity fields used were taken from European Centre for Medium-Range Weather Forecasting (ECMWF) initialized operational data from the period November 1996 to May 1998, which is available at 6-hour intervals. The wind data at T42 horizontal resolution was interpolated to the vertical isentropic surfaces, and the nondivergent part was extracted. Linear interpolation in time within each 6-hour interval was then used to provide a continuously varying time series of velocity field as input for the model. For the simulations to be reported the spectral model resolution was taken to be T83, thereby allowing the tracer to develop finer-scale structure than was present in the velocity fields. Unless otherwise stated, the diffusivity \( \kappa \) was taken to be \( 3.24 \times 10^7 \text{ m}^2 \text{ s}^{-1} \), corresponding to a damping time of 0.2 days for the smallest resolved scale. We believe that the “memory time” implied by this is sufficiently small to allow \( \kappa_{\text{eff}} \) to resolve variations in the mixing ability over timescales of about a week.

The tracer field was output from the simulation at daily intervals, and for each output the effective diffusivity \( \kappa_{\text{eff}} \) was calculated as a function of \( \phi_e \). The method of calculating \( \kappa_{\text{eff}} \) is as follows. The gradient of the tracer is calculated at each grid point, and its square is integrated over the area bounded by the desired tracer contour. This integrated square gradient is then differentiated with respect to area by taking finite differences. The resulting quantity is then divided by the square of the areal gradient of the tracer, and multiplied by \( \kappa \) to obtain \( \kappa_{\text{eff}} \) as defined in equation (3).

2.3. Dependence on Initial Tracer Field

The novel aspect of the investigation reported here is that the properties of the flow are diagnosed from the effective diffusivity calculated, as a function of equivalent latitude, from a single advected tracer field. It is therefore necessary to estimate that the effective diffusivity is determined primarily by the flow and is not strongly dependent on the initial condition imposed upon the tracer field.

It is useful to consider steady two-dimensional flow as a prototype, since it highlights some of the behavior of effective diffusivity that may be expected in more general flows. In a steady two-dimensional flow, for small values of the diffusivity, there is a well understood two-stage process involved in the evolution of a tracer field from a given initial condition [Rhines and Young, 1983]. In the first, rapid, stage the tracer field is homogenized along the streamlines of the steady flow. The time required for this first stage increases as the diffusivity decreases. In the second, slower, stage there is diffusion across the streamlines (and continuous rapid rehomogenization of the tracer along the streamlines). Since contours of tracer and streamlines are almost exactly aligned in this second stage, a coordinate based on streamlines is almost-exactly equivalent to one based on area inside tracer contours (or equivalent latitude). Rhines and Young [1983] give an explicit expression for the diffusivity across each streamline in this second stage, valid in the limit of small diffusivity. Nakamura and Ma [1997] remark that, in the same limit, this cross-streamline diffusivity is equal to the corresponding effective diffusivity (with the implicit assumption that there is a unique correspondence between tracer concentration values and streamlines). Since the Rhines and Young diffusivity is defined without reference to any particular tracer field, it follows that in this case the effective diffusivity can, in the second stage of the evolution, indeed be regarded as a diagnostic of the flow.

In a time-dependent flow, even in the small-diffusivity limit, there will be no almost-exact alignment of tracer contours with streamlines, and the above argument does not hold precisely. However, we expect for the realistic atmospheric flows to be considered that transport in the latitudinal direction will be, broadly speaking, slower than transport in the longitudinal direction. It is therefore to be expected that tracer will, in a first stage, adjust to align with the dominant quasi-longitudinal flow and that subsequently it will, in a second stage, adjust more slowly in latitude. The first stage will require diffusive decay of the structure that arises purely from the initial condition, and the time required will therefore depend on the diffusivity. By analogy with the steady flow case above, we expect the effective diffusivity to be approximately independent of the initial condition after a first stage of adjustment (provided, also by analogy with above,
there is a unique correspondence between initial tracer and latitude). The unique correspondence, which requires that the tracer is a monotonic function of latitude, also ensures that the equivalent latitude coordinate $\phi_e$, defined with reference to the tracer, is a useful measure of actual latitude $\phi$.

We first examine whether the tracer indeed adjusts to a configuration for which the effective diffusivity is independent of the initial conditions, as argued above. We choose the initial tracer to be $\sin(\phi)$, as a simple initial condition that satisfies the monotonicity requirement. Two simulations were carried out using winds for the 450 K isentropic surface, the first beginning November 1, 1997, and the second beginning December 1, 1997, and both continuing until February 28, 1998. Figure 2a shows the effective diffusivity as a function of equivalent latitude for the tracer from the two simulations on December 6, January 1, 1998, and February 28, 1998. A 10-day running-mean was applied to the effective diffusivity (as for the time evolution of effective
Figure 3. (a) Five different tracer initial conditions shown by different curves: solid, $\mu$; dot, $\phi$; dash, $\mu^2$; dash-dot, $\tanh(2\phi)$; and dash-dot-dot-dot-dot, $\mu - \frac{5}{3}(\mu/\sqrt{2})^3 + \frac{2}{3}(\mu/\sqrt{2})^5$. (b) The $\kappa_{\text{eff}}(\phi_e)$ distribution on December 1, 1997, from five simulations using winds from 450 K, beginning November 1, 1997, with the tracer initial conditions shown in (a).

Diffusivity to be shown later in Plate 4), and then the percentage difference (averaged over all equivalent latitudes) in the effective diffusivity between the two simulations was calculated. The time evolution of this percentage difference is shown in Figure 2b. The effective diffusivity distribution in the two simulations are already surprisingly similar on December 6, 1997, after only 6 days of the second simulation. The distributions on January 1, 1998 and February 28, 1998, after 1 and 3 months, respectively, of the second simulation, show much closer agreement, and the time evolution shown in Figure 2b confirms that after January 1 there is never more than about a 5% difference between the two simulations. (The similarity of the two distributions on February 28 also demonstrates that $\kappa_{\text{eff}}$ calculated in this manner remains a useful diagnostic for at least 3 months.)

The first simulation above, which began November 1, 1997, was repeated with different initial conditions for the tracer field. The different initial conditions were chosen
to give high and low gradients in different regions and are shown in Figure 3a together with, in Figure 3b, the corresponding effective diffusivities on December 1, 1997, after 1 month of simulation. The distributions are essentially the same, the differences appearing only in small-scale features. Such small-scale features will not be regarded as significant in the discussion that follows in the remainder of the paper.

On the basis of these results we argue that after a time of about 1 month the effective diffusivity is independent of the tracer initial condition and therefore that after this time it may indeed be regarded as a diagnostic of the properties of the flow itself. Consequently, for all the simulations reported later in the paper, a "spin-up" time of 1 month is allowed after the beginning of the simulation before effective diffusivity is first calculated.

2.4. Quantitative Relation Between Effective Diffusivity and Flow Properties

It has been demonstrated by Nakamura [1996], and argued above, that low values of effective diffusivity correspond to barrier regions and high values to mixing regions. The precise relation between effective diffusivity and other more standard measures of transport and mixing will be discussed in detail in a future paper, but one potentially unattractive feature is the apparent dependence, explicit in definition (3), of effective diffusivity on the diffusivity $\kappa$.

We argue below that, in the approach taken in this paper, $\kappa$ can essentially be chosen arbitrarily, within certain numerical constraints (one being that it should be large enough to prevent strong spatial variation of the tracer on the scale of the grid). Certainly there is no "correct" value of $\kappa$, and if the effective diffusivity is to be a precise quantitative measure of the transport and mixing ability of a flow, then it should surely be independent of $\kappa$, at least in some suitable limit. Nakamura [1996] speculates that $\kappa_{\text{eff}}$ may tend to a finite value in the limit $\kappa \to 0$, and it is possible to construct heuristic arguments that support this idea for the flow in mixing regions. However, this independence of $\kappa$ cannot be a universal property. Consider the case when there are perfect, absolutely impermeable, barriers to transport (as is often the case in time periodic chaotic-advection flows, for example). Then tracer contours will align with those barriers and will consequently have the same geometric structure. The geometric structure of the barriers is independent of $\kappa$, and it follows that $L^2_{\text{eff}}$ will remain finite in the limit as $\kappa \to 0$. This implies, from (3), that $\kappa_{\text{eff}}$ will be proportional to $\kappa$ in this limit.

Similar issues arise in calculating and interpreting $\kappa_{\text{eff}}$ from observations with limited spatial resolution, and are discussed by Nakamura and Ma [1997]. Here we follow Nakamura and Ma in assuming that, if the structure, but not necessarily the magnitude, of $\kappa_{\text{eff}}$ is largely independent of $\kappa$, then we can extract useful information about the shape and location of barriers. We also assume that it is valid to compare the relative strength of barriers by considering the relative magnitudes of $\kappa_{\text{eff}}$, for the same value of $\kappa$, i.e., that for given $\kappa$, larger $\kappa_{\text{eff}}$ implies relatively stronger mixing and relatively weaker barriers.

We now present results of selected simulations at different values of $\kappa$ to support the idea that the structure of $\kappa_{\text{eff}}$ is to good approximation independent of $\kappa$. Each simulation was at 185 resolution, used wind fields for the 450 K isentropic surface, and covered the period December 1, 1997 to February 28, 1998. The values of diffusivity $\kappa$ were taken to be $1.62 \times 10^3 \text{ m}^2 \text{ s}^{-1}$, $3.24 \times 10^3 \text{ m}^2 \text{ s}^{-1}$, and $6.48 \times 10^3 \text{ m}^2 \text{ s}^{-1}$, corresponding to damping times for the smallest resolved scale of 0.4, 0.2, and 0.1 days, respectively. We compared the effective diffusivity distribution between the different simulations during the period January 1, 1998 to February 28, 1998, i.e., after the chosen spin-up time of 1 month. Figure 4 shows the variation of the effective diffusivity with equivalent latitude in the three simulations for one particular day, February 19, 1998. For each of the three simulations, the effective diffusivity shows two maxima corresponding to mixing regions, the larger in the Southern Hemisphere midlatitudes and the smaller in the Northern Hemisphere midlatitudes. In each case there are also two minima, corresponding to barrier regions, the first in Northern Hemisphere high latitudes and the second a broader minimum in low latitudes, both with very similar values. The fact that these properties hold over all three simulations strongly supports the idea that the effective diffusivity may be used to identify, and to make quantitative comparison between, transport and mixing regions.

The effective diffusivity also shows some of the expected variation with $\kappa$, discussed earlier. In the barrier regions the values of $\kappa_{\text{eff}}$ appear approximately proportional to $\kappa$. In the mixing regions the values of $\kappa_{\text{eff}}$ appear less dependent on $\kappa$, particularly in the stronger mixing region in the Southern Hemisphere.

In the following we assemble a time series of effective diffusivity by performing a series of 4-month integrations for each isentropic surface with the following potential temperature $\theta$: 400, 450, 500, 550, 600, 700 and 850 K. Only the results from the last 3 months of each integration were used (to allow the spin-up time mentioned earlier). For each integration the tracer was initialized with $\mathbf{v} = \mathbf{0}$. In fact, such re-initialization is probably unnecessary over timescales of a year or so, while useful contrast remains in the tracer field. On longer timescales there may be a problem that the tracer homogenizes, but this can be avoided using a weak source term, as will be discussed in a future paper. The results will be presented in the form of effective diffusivity divided by the diffusivity (i.e., $\kappa_{\text{eff}} = \kappa_{\text{eff}} / \kappa$). This is a nondimensionalized quantity that takes the value 1 when a tracer contour is geometrically equivalent to a latitude circle.

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The effective diffusivity for all equivalent latitudes on isentropic surfaces in the range 400–850 K is shown as monthly averages for the period June 1997 to November 1997 (Southern Hemisphere winter/spring) in Plate 1 and for the period December 1997 to May 1998 (Northern Hemisphere winter/spring) in Plate 2. The zonal-mean zonal winds are overlaid, plotted against latitude rather than equiv
Figure 4. The $\kappa_{\text{eff}}(\phi_e)$ distribution on February 19, 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 450 K surface and using an initial condition of $\sin(\phi)$ for the tracer on December 1, 1997.

alent latitude; this will give a rough idea of the associated flow structure, although it clearly has limitations.

The zonal winds show a strong westerly jet, the polar vortex, in the Southern Hemisphere, centered at about 60° for the months of April and May (Plate 2) and June to November (Plate 1). Centered on the core of the jet is a clear band of low $\kappa_{\text{eff}}$ (i.e., $\ln \kappa_{\text{eff}} < 1$). A similar band of low $\kappa_{\text{eff}}$ is visible centered on the core of the Northern Hemisphere polar vortex in the months November (Plate 1) and December to March (Plate 2). These bands of low $\kappa_{\text{eff}}$ indicate barriers to transport that will partially isolate the air within the polar vortex from that outside. We shall refer to these as “vortex-edge barriers.” Also seen in the winter hemispheres, equatorward of the vortex-edge barriers, are broad bands of high $\kappa_{\text{eff}}$ (i.e., $\ln \kappa_{\text{eff}} > 2$). These midlatitude high values of $\kappa_{\text{eff}}$ indicate strong mixing and correspond to the so-called “surf zone.” Hence the vortex/surf-zone structure, first noted by McIntyre and Palmer [1983], is clearly evident in both hemispheres in the effective diffusivity. The transport and mixing associated with this vortex/surf-zone structure varies as the winter and spring seasons progress.

We now consider the seasonal evolution in detail. We will first consider the Southern Hemisphere winter (June 1997 to November 1997) and then the Northern Hemisphere winter (December 1997 to May 1998), and note the differences between them. We then compare this Northern Hemisphere winter with the previous winter (December 1996 to May 1997).

3.1. Southern Hemisphere Winter: Antarctic Vortex

The vortex-edge barrier of the Antarctic polar vortex is clearly apparent in the middle stratosphere (600–850 K) in June as a region of low $\kappa_{\text{eff}}$ ($\ln \kappa_{\text{eff}} < 1$) associated with the core of the jet between 40°S and 70°S (Plate 1). There are slightly higher $\kappa_{\text{eff}}$ values ($1 < \ln \kappa_{\text{eff}} < 1.5$) equatorward of the jet maximum (in the surf zone) and poleward of it (within the vortex), indicating some weak mixing in these regions. In the lower stratosphere (450–600 K) the entire region associated with the vortex and its interior is characterized by low $\kappa_{\text{eff}}$, with the lowest $\kappa_{\text{eff}}$ (in $\ln \kappa_{\text{eff}} < 0.5$) centered slightly poleward of the jet core. At these levels there is apparently strong mixing ($1.5 < \ln \kappa_{\text{eff}} < 2$) outside the vortex.

As the zonal winds strengthen at lower levels in the subsequent months (July and August), there are low values of $\kappa_{\text{eff}}$ ($\ln \kappa_{\text{eff}} < 1$) down to 400 K, with the lowest values of $\kappa_{\text{eff}}$ on each level coincident with the maximum zonal wind at about 60°. During these months the values of $\kappa_{\text{eff}}$ in the vortex edge are very low ($\ln \kappa_{\text{eff}} < 0.25$), suggesting that the vortex edge acts as a very effective barrier, particularly in the lower stratosphere where minimum values of $\kappa_{\text{eff}}$ are lowest and the region of low $\kappa_{\text{eff}}$ is broad. There continues to be a distinction between the middle and the lower stratosphere in the sense that there are significantly higher values of $\kappa_{\text{eff}}$ within the vortex at upper levels ($1 < \ln \kappa_{\text{eff}} < 1.5$) than at lower levels ($\ln \kappa_{\text{eff}} < 1$).

In September and October there are low values of $\kappa_{\text{eff}}$ in the entire region within the vortex, indicating little mixing
Plate 1. The $\tilde{\kappa}_{\text{eff}}(\phi_e, \theta)$ distribution, averaged over each month during the period June 1997 to November 1997, and superimposed, contours of zonal-mean zonal wind (as a function of $\phi_e$ and $\theta$), with easterlies dashed and westerlies solid.
Plate 2. The $\tilde{\kappa}(\phi, \theta)$ distribution, averaged over each month during the period December 1997 to May 1998, and superimposed, contours of zonal-mean zonal wind (as a function of $\phi$, and $\theta$), with easterlies dashed and westerlies solid.
Plate 3. The $\tilde{\kappa}_{\text{eff}}(\phi_e, \theta)$ distribution, averaged over each month during the period December 1996 to May 1997, and superimposed, contours of zonal-mean zonal wind (as a function of $\phi_e$ and $\theta$), with easterlies dashed and westerlies solid.
Plate 4. The $\kappa_{\text{eff}}(\phi, t)$ distribution for December 1996 to May 1998 on the 400, 450, 550, and 850 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.
there. Values of $\kappa_{\text{eff}}$ in the vortex edge remain very low in the lower stratosphere, though the width of the region of low values has decreased from July and August. Of note is that in the middle stratosphere the edge of the region of low $\kappa_{\text{eff}}$ has moved poleward by about 10°, from 50°S in August to 60°S in September. In October the values of $\kappa_{\text{eff}}$ in this region have increased slightly (ln $\kappa_{\text{eff}}$ > 0.25). These two effects (the decrease in size of the vortex and the weakening of the barrier of its edge) are consistent with the idea of vortex erosion, as filaments are stripped from the vortex edge.

In June and July values of $\kappa_{\text{eff}}$ in the midlatitude surf zone are rather low ($1 < \ln \kappa_{\text{eff}} < 2$), indicating weak mixing, particularly in the middle stratosphere. In August there is apparently a dramatic strengthening of the surf-zone mixing indicated by high values of $\kappa_{\text{eff}}$ (ln $\kappa_{\text{eff}} > 2$), particularly above 700 K. The width of the surf zone is also substantially increased. These changes are presumably as a result of an increase in planetary-wave activity from July to August. While the geopotential amplitudes of stationary wavenumber 1 are similar in August to those in June and July, the amplitudes of transient waves, particularly wavenumber 2, are much increased [Randel, 1992], and this may account for the increase in the strength of mixing in the width of the surf zone. The surf-zone mixing remains strong in September and October and presumably contributes to the vortex erosion mentioned above.

Throughout the winter, values of $\kappa_{\text{eff}}$ in the surf zone are higher than those inside the vortex (except above 700 K in June). This is consistent with Schoeberl et al. [1992] who found eddy mixing rates in the vortex to be an order of magnitude less than in the surf zone for spring 1987.

In November there is a significant change. The zonal winds show only a weak vortex remaining at low levels. Above about 600 K the high values of $\kappa_{\text{eff}}$, although maximal in midlatitudes, extend to the pole, and there is no sign of a vortex-edge barrier. The high values of $\kappa_{\text{eff}}$ presumably indicate the mixing processes associated with the breakup of the vortex. Below about 550 K there are some remnants of the vortex-edge barrier structure, indicated by moderately low values of $\kappa_{\text{eff}}$.

3.2. Northern Hemisphere Winter: Arctic Vortex

The Arctic vortex-edge barrier is apparent in the middle stratosphere (600–850 K) in October as a region of low $\kappa_{\text{eff}}$ (ln $\kappa_{\text{eff}} < 1$) poleward of about 50°N (Plate 1). There are moderate values of $\kappa_{\text{eff}}$ (1 < ln $\kappa_{\text{eff}} < 1.5$) equatorward of the barrier, where the surf zone is developing.

In November the surf zone is broader and the mixing is stronger, as indicated by moderately high values of $\kappa_{\text{eff}}$ (1.5 < ln $\kappa_{\text{eff}} < 2$). In the lower stratosphere (450–600 K) there is a narrow band of quite low $\kappa_{\text{eff}}$ (0.5 < ln $\kappa_{\text{eff}} < 1$) located between 60°N and 75°N, with higher values on both the equatorward and the poleward sides. The picture of this period gained from the $\kappa_{\text{eff}}$ distribution, of a vortex edge barrier that is reasonably well developed at high levels but much weaker at lower levels, is consistent with other measures, such as wind, temperature, and ozone [Third European Stratospheric Experiment on Ozone (THESEO), 1998].

December shows weaker zonal winds (Plate 2). This may be explained by the series of warming events in December (December 8–13 and December 22 to January 8) which significantly eroded but did not break up the vortex, as documented in the Third European Stratospheric Experiment on Ozone (THESEO) [1998]. In the middle stratosphere the low values of $\kappa_{\text{eff}}$ are confined to a narrow band between 60°N and 75°N at this time (Plate 2), indicating only a rather weak vortex-edge barrier. In the lower stratosphere (450–600 K) the barrier is weaker still, and below this there are moderate values of $\kappa_{\text{eff}}$ (1 < ln $\kappa_{\text{eff}} < 1.5$) and no vortex-edge barrier. The vertical variation appears to be, in the monthly average, a gradual weakening through the "sub-vortex" transition. Examination of daily effective diffusivity distributions indicate that although this transition point descends through the winter season, thereby smoothing the monthly averages, there is always a gradual transition rather than the sharp one that has been suggested by McIntyre [1995]. The "sub-vortex" issue will be discussed further in part 2. The interior of the vortex is characterized by moderate values of $\kappa_{\text{eff}}$, indicating weak mixing.

The zonal winds are weaker still in January, indeed the vortex has almost disappeared. Of course, this signature in the zonal-mean zonal winds (with the mean taken at fixed latitude) may be due to dynamical effects that are largely reversible, such as displacement of the vortex away from the pole. By construction, $\kappa_{\text{eff}}$, which is a function of equivalent latitude rather than latitude, is expected to emphasize irreversible dynamical changes. Indeed, in contrast to the zonal winds, the $\kappa_{\text{eff}}$ distribution for January 1998 is generally similar to that for December, with a clear minimum in $\kappa_{\text{eff}}$ in the middle stratosphere indicating that a fairly strong vortex-edge barrier persists. The small changes from December to January are that the lower limit of the vortex-edge barrier has risen to 550 K and that there is no sign of mixing within the vortex.

There is a significant change in the $\kappa_{\text{eff}}$ distribution in February. At all levels in the stratosphere (450–850 K) there are very low values of $\kappa_{\text{eff}}$ (ln $\kappa_{\text{eff}} < 0.5$) poleward of 65°, indicating the vortex edge acts as a very strong barrier at this time. This includes the lower levels (450–600 K) where the vortex-edge barrier was weak or nonexistent in earlier months. A deep vortex structure is clearly visible in the zonal winds, consistent with the $\kappa_{\text{eff}}$ distribution and, incidentally, with the cold, sub PSC threshold, temperatures observed for part of this time [Third European Stratospheric Experiment on Ozone (THESEO), 1998]. The mixing in the surf zone is seen to have reduced in strength, with moderate values of $\kappa_{\text{eff}}$ (1 < ln $\kappa_{\text{eff}} < 2$).

In March the zonal winds are seen to have weakened considerably and the high-latitude $\kappa_{\text{eff}}$ to have increased (suggesting that this change in the vortex is not purely reversible). By April 1998 all trace of a high-latitude vortex in the zonal winds has disappeared, and there are high values of $\kappa_{\text{eff}}$ throughout the region poleward of about 45°N, presumably associated with the final breakup of the vortex. There is a hint of a weak jet and a corresponding transport barrier at about 45°N above 700 K. This is reminiscent of
the Northern Hemisphere final warming of 1979 described by Hess [1991], in which a westerly jet at midlatitudes and corresponding relatively large gradients in ozone and potential vorticity appear to persist through to mid-April after a warming event at the end of February.

From April to May the values of \( k_{\text{eff}} \) reduce substantially in the middle stratosphere, indicating a transition in the Northern Hemisphere to the summer regime (to be discussed further in section 3.4).

In the Southern Hemisphere the development of the Antarctic vortex is visible in the winds and in the \( k_{\text{eff}} \) distribution through the months March, April, and May. The high-latitude region of moderate values of \( k_{\text{eff}} \) in the middle stratosphere in May indicates mixing within the vortex, perhaps associated with deformation of the vortex itself.

We now compare the winter and spring seasons in Northern (Plate 2) and Southern (Plate 1) Hemispheres.

December 1997 and January 1998 look very different both in surf-zone and vortex structure compared with June and July 1997. The Arctic vortex-edge barrier is narrower (15° in the middle stratosphere) and weaker (0.5 < ln \( k_{\text{eff}} \) < 1) than the Antarctic vortex-edge barrier (30° and ln \( k_{\text{eff}} \) < 0.5). The differences in strength are particularly striking in the lower stratosphere (450–600 K), where the values of \( k_{\text{eff}} \) associated with the vortex edge and the vortex interior are considerably larger in the Arctic (1 < ln \( k_{\text{eff}} \) < 1.5) than in the Antarctic (ln \( k_{\text{eff}} \) < 0.5). The region enclosed by the Arctic vortex-edge barrier is smaller, with the edge defined by the lowest values of \( k_{\text{eff}} \) (60°N at 830 K) almost 20° poleward of that in the Antarctic (40°S at 850 K). There is a broad region of very high values of \( k_{\text{eff}} \) (ln \( k_{\text{eff}} \) > 2) in the surf zone of the Northern Hemisphere (December and January), whereas there is a narrow region of moderate values (1 < ln \( k_{\text{eff}} \) < 1.5) in the Southern Hemisphere (June and July).

The strengths and widths of the vortex-edge barriers are similar in the Northern Hemisphere in February 1998 and in the Southern Hemisphere in August 1997, since the width of the Antarctic barrier reduces from July to August and the strength of the Arctic vortex increases from January to February. (It seems that vortex erosion at this stage of this particular winter is much more effective in the Southern Hemisphere than in the Northern Hemisphere.) March and April 1998 are clearly different to September and October 1997, with little or no barrier in the Northern Hemisphere.

Mixing in the interior of the vortex, which appears to be confined to the midwinter middle stratosphere, is stronger in the Southern Hemisphere (1 < ln \( k_{\text{eff}} \) < 1.5) than in the Northern Hemisphere (0.5 < ln \( k_{\text{eff}} \) < 1).

During the first half of the winter the surf-zone mixing is strong (ln \( k_{\text{eff}} \) > 2) in the Northern Hemisphere (December 1997 and January 1998), but much weaker (1 < ln \( k_{\text{eff}} \) < 2) in the Southern Hemisphere (June and July 1997). But, perhaps surprisingly, the situation is reversed in the second half of the winter (February and March versus August and September).

### 3.3. Northern Hemisphere Winter 1996/1997

There is known to be substantial interannual variability in the dynamical behavior of the Northern Hemisphere winter stratosphere. To give some indication of the range of transport and mixing behavior that is possible in Northern Hemisphere winter, effective diffusivity was also calculated for the period December 1996 to May 1997 and results are shown in Plate 3. Again, the zonal-mean zonal winds are overlaid (plotted against latitude rather than equivalent latitude) to give a rough idea of the associated flow structure. The circulation for this period has been described by Coy et al. [1997].

In December 1996 there is no strong vortex visible from the zonal winds; however, the values of \( k_{\text{eff}} \) are low at latitudes greater than about 50°N, at least above 600 K. Below this, values of \( k_{\text{eff}} \) are higher and any vortex-edge barrier must be very weak. At this early stage of the winter, the vortex and the vortex-edge barrier were less well developed at lower levels, both in terms of stronger zonal winds and lower values of \( k_{\text{eff}} \), than in December 1997.

The region of low \( k_{\text{eff}} \) associated with the vortex-edge barrier in January is seen to be broader and deeper in 1997 (extending to 65°N and 450 K) than in 1998 (extending to 60°N and 550 K), and correspondingly, the surf-zone mixing is generally weaker, particularly at lower levels. However, the Northern Hemisphere winter of 1996/1997 cannot simply be characterized as less disturbed than that of 1997/1998 since the vortex-edge barrier in February 1997 is seen to be of comparable strength to that in February 1998.

There are significant differences between 1997 and 1998 in late winter. In particular, a strong vortex-edge barrier persists through March and April 1997, and the strong mixing associated with the breakdown of the vortex is not seen until May.

### 3.4. Summer Hemisphere

The summer hemispheres, both Northern (Plate 1, June, July, and August) and Southern (Plates 2 and 3, January and February) are seen to be characterized by high values of \( k_{\text{eff}} \), i.e., strong mixing, in the lower stratosphere, in particular at the lowest levels 400–500 K. The values of \( k_{\text{eff}} \) decrease in the vertical and are very low in the middle stratosphere. It is likely that both baroclinic eddies, whose effects penetrate upward from the tropopause region [Miles and Gleason, 1986], and breaking planetary-scale waves propagating on the weak westerlies, make a contribution to the mixing in the lower stratosphere. Note that there appears to be a strong association between the regions where \( k_{\text{eff}} \) takes high values and regions where the zonal wind are westerly (even only weakly so). This is particularly clear in the transition from January 1997 to March 1997 in the Southern Hemisphere, seen in Plate 3. Both the region where winds are westerly and the region of higher values of \( k_{\text{eff}} \) are deeper in February than in January. Then in March the region of high values of \( k_{\text{eff}} \) is deeper still, as what appears to be the beginning of a surf
one forms in a deep region centered at about 50°S. This suggests that the upward propagation and reemergence of planetary waves which is allowed by the weak westerlies.

Note also that while June 1997 in the Northern Hemisphere (Plate 1) has a clear summer distribution of $\kappa_{\text{eff}}$, December 1996 and December 1997 in the Southern Hemisphere (Plates 2 and 3) are very different, with high values of $\kappa_{\text{eff}}$ extending up to 850 K. This is almost certainly explained by the later breakup of the vortex in the Southern Hemisphere, with the high values of $\kappa_{\text{eff}}$ in December associated with the vortex breakup itself.

The clear evidence of strong mixing in the lower stratosphere of the summer hemisphere is perhaps contrary to the standard view of the summer stratosphere as relatively quiescent. However, it is consistent with particle trajectory studies reported by Bowman [1996] and Sparling et al. [1997] (both for the 500 K level in the Southern Hemisphere summer). It is also consistent with observed transport of aerosol from tropical volcanic eruptions, for example, the transport of Pinatubo aerosol into the Northern Hemisphere extratropics during the period June–September 1991 reported by Repte et al. [1993].

The entire seasonal evolution in the extratropics is particularly clear from Plate 4, which shows $\kappa_{\text{eff}}$ as a function of equivalent latitude and time (covering the entire period from December 1996 to May 1998) on four isentropic surfaces (400, 450, 550, and 850 K). At 400 K there is a weak, seasonal variation of asymmetry between the hemispheres, consistent with the dominant mixing coming from synoptic-scale eddies. As height increases, the contrast between winter and summer becomes more pronounced with the increasing role of synoptic-scale eddies and the increasing importance of stratospheric planetary waves. The effect of inhibitor mixing at the vortex-edge barrier becomes weaker. The upper limit of the mixing in the summer hemisphere is close to 550 K, so that at 850 K it has disappeared altogether.

5. Tropics

It is now widely accepted that the tropical lower stratosphere acts as a reservoir for chemical species and volcanic aerosol and that it therefore must be relatively isolated from extratropical transport by stratospheric transport barriers. Chemical species with a tropospheric source are transported vertically through the reservoir, in a “tropical pipe” [Plumb, 1996] and slowly into the extratropics. There have been a number of studies assessing the rate of isentropic transport across the subtropical barrier, both outward, from the tropical reservoir into the extratropics, and inward. The more recent studies have considered the vertical variation of this rate of isentropic transport and suggest that it is smallest (i.e., the tropics are most isolated) at around 550 K, on the basis of chemical observations [Mote et al., 1998, Münchauer et al., 1996], volcanic aerosol [Grant et al., 1996], material contour advection [Waugh, 1996] and chemical transport model simulations [Chen et al., 1994a; Rogers et al., 1999].

Plates 1, 2, and 3 show that values of $\kappa_{\text{eff}}$ in the tropics are generally very low, indicating very weak transport into and out of this region. Thus the distribution of $\kappa_{\text{eff}}$ is consistent with the idea of a tropical reservoir. If subtropical transport barriers are to be defined, it seems best to associate them with the subtropical boundaries of the low-latitude region of low $\kappa_{\text{eff}}$. Note that there is no subtropical minimum of $\kappa_{\text{eff}}$ that may be used to identify a subtropical barrier in the way that a high-latitude minimum may be used to identify the vortex-edge barrier. This point has been noted previously with respect to stretching rates by Chen [1996].

The first three panels of Plate 4 (400, 450, and 550 K) show a pronounced broadening with height of the tropical band of low $\kappa_{\text{eff}}$. At the same time, there is also a seasonal cycle which changes its structure significantly with height.

At the lowest levels, exemplified by 400 K, the tropical reservoir covers a region of width between 20° and 40° depending on the season. The boundaries of the reservoir mark the transition to the region dominated by the strong mixing effects of synoptic eddies, which break on the upper equatorward flank of the subtropical jet. These boundaries move poleward to give a maximum width of the tropical reservoir in August–October and equatorward to give a minimum width in February–March. In the Northern Hemisphere, the late summer (August–October) poleward displacement of the boundary could be associated with the summer poleward displacement of the Northern Hemisphere subtropical jet. In the Southern Hemisphere, the subtropical jet also displaces poleward in summer, but the winds remain westerly at relatively low latitudes, and this may account for the lack of poleward displacement of the boundary in the Southern Hemisphere in late summer (February–March). The broadening of the tropical reservoir in August–October is also associated with the weakened mixing in the subtropics in the Southern Hemisphere during these spring months. There is an analogous weakening in Northern Hemisphere spring, but it takes a slightly different form and does not seem to affect the boundary of the tropical reservoir.

At higher levels in the lower stratosphere, such as 450 K and 550 K, the differences between winter and summer become stronger. However, at these levels, the seasonal cycle takes the form of a shift toward the summer pole (about 10° at 450 K and about 70° at 550 K) of the entire tropical reservoir, while the width remains of roughly constant throughout the year (45° at 450 K and 55° at 550 K). There are high values of $\kappa_{\text{eff}}$ in the summer subtropics, presumably associated primarily with mixing by synoptic-scale eddies, and high values in the winter subtropics, presumably associated primarily with breaking planetary waves. The latter penetrate farther equatorward than the former and hence the tropical reservoir is shifted toward the summer hemisphere.

A similar seasonal cycle in the position of the tropical reservoir has also been noted in observations, for example, from satellite data of aerosol following the Mount Pinatubo eruption [Grant et al., 1996, Figure 3], and total ozone data [Grant et al., 1996, Figure 6]. High tropical values of methane at 68 hPa (~450 K) are also observed to shift
towards the summer pole [Randel et al., 1998, Figure 9]. However, the seasonal cycle in observed tracers may be influenced by factors other than isentropic transport. For instance, the diabatic upwelling also shows a seasonal shift in position [Randel et al., 1998; Plumb and Eluszkiewicz, 1999]. Indeed, an interesting coincidence is that the seasonal shift in upwelling closely matches that in effective diffusivity, though it is difficult to see why there should be any correspondence between the two.

Waugh [1996] suggested that there is a seasonal variation in the transport out of the tropics. He found, at 500 K, less transport during the summer months, suggesting a stronger barrier, than during other seasons (see [Waugh, 1996, Figures 5 and 15]). However, as shown by Plate 4, the seasonal variation in transport may actually result from the seasonal shift in position of the tropical reservoir, with the region of high $\kappa_{\text{eff}}$ being further from Waugh's 20° reference latitude in summer, rather than from a seasonal variation in strength of the subtropical barrier.

The broadening with height of the tropical reservoir between 400 K and 600 K, seen in Plate 4, may complicate the interpretation of previous calculations of transport from the tropics. For example, Waugh [1996] considered material contours initialized at 20° and calculated the area starting equatorward of such contours that was transported north of 30°, while Chen et al. [1994a] considered the evolution of tracer initialized to a strip of width 20° either side of the equator. Both concluded that the subtropical transport barrier was stronger at 550 K than at 400 K or 450 K. However, the results of Waugh and Chen et al. could equally well be reinterpreted as a consequence of the broadening with height of the tropical reservoir, since parts of a material contour initialized at 20° are less likely to move poleward at 550 K than at the lower levels (particularly in summer).

The final panel of Plate 4 shows that in the middle stratosphere, at levels such as 850 K, the seasonal variation becomes stronger. In a sense, the shift in position of the tropical reservoir into the summer hemisphere is so great that low values of $\kappa_{\text{eff}}$ extend all the way to the pole. One effect is that at high latitudes there is a sharp transition between high values of $\kappa_{\text{eff}}$ in spring as the vortex breaks up and very low values in summer. The limit of the tropical reservoir is well defined on the winter side by the subtropical boundary of the surf zone. This boundary often intrudes to latitudes less than 20°. This variation seems consistent with the seasonal variation in methane at 10 hPa (~850 K) where relatively high tropical values mix out into midlatitudes in winter [Randel et al., 1998, Figure 9]. As a result of the stronger seasonal variation at this level, the width and isolation of the tropical reservoir defined in an annual average sense is less than at 550 K, as is consistent with previous studies.

3.6. Interannual Variability: Subtropics and Tropics

Inspection of Plates 1 to 4 suggests a clear relation between variations in zonal winds in the tropics associated with the quasi-biennial oscillation (QBO) and variations in the effective diffusivity.

QBO westerlies (Plates 1 and 3) first appear at 850 K in February 1997 and by May 1997 have descended to about 450 K. The westerlies are seen to be narrow, as expected from the standard picture of the QBO. Within these narrow westerlies the values of $\kappa_{\text{eff}}$ are low (ln $\kappa_{\text{eff}} < 1$ before May 1997 and ln $\kappa_{\text{eff}} < 0.5$ from then until October 1997). During winter/summer months, in the subtropics on the summer hemisphere side of these westerlies there is a region of higher $\kappa_{\text{eff}}$ (1 < ln $\kappa_{\text{eff}} < 1.5$ before April 1997 and 0.5 < ln $\kappa_{\text{eff}} < 1$ between June 1997 and October 1997). During spring/autumn months, on the flanks of the westerlies, in both hemispheres, the values of $\kappa_{\text{eff}}$ are higher (1 < ln $\kappa_{\text{eff}} < 1.5$ in April and May 1997). Looking at 850 K during the period March 1997 to June 1997 (Plate 4), it is interesting to note that the net effect of the QBO and the seasonal evolution is that the region of westerly winds moves southward to the equator, and that the corresponding region of higher $\kappa_{\text{eff}}$, initially in the summer subtropics to the south of the westerlies, moves poleward (with the region of weak winds) as the tropical westerlies increase in strength.

The QBO easterlies present at other times (Plates 1, 2, and 3) are, again as expected, broad and within these broad easterlies the values of $\kappa_{\text{eff}}$ are generally very low. There are, however, slightly higher values of $\kappa_{\text{eff}}$ in the northern tropics and subtropics in December 1996 and January 1997 than in December 1997 and January 1998 (see 850 K in Plate 4). The tropical winds are easterly in both winters, but they are weak in the first winter and the extratropical westerlies reach closer to the equator. This should allow extratropical planetary waves to propagate to lower latitudes in the first winter, and it is consistent with the higher values of $\kappa_{\text{eff}}$ seen in that winter.

These results strongly suggest that, as in the model simulations of Polvani et al. [1995] and O'Sullivan and Chen [1996] (see also Chen [1996]), when there are easterly QBO winds, waves are inhibited from penetrating into the tropics and hence tropical mixing is weak. Conversely, when there are westerly QBO winds, waves are able to propagate across the equator and break, causing mixing, in the weaker winds on the summer hemisphere side of the westerlies (see, in particular, O'Sullivan and Chen [1996]).

The differences we find in $\kappa_{\text{eff}}$ between easterly and westerly winds indicate clearly that the QBO has a strong effect on isentropic mixing over a substantial range of levels. This appears to disagree with the conclusions of Waugh [1996] (although it may reflect the potential complications, noted earlier, in interpreting his results). It is also the case that Gray and Russell [1999] have argued that, in the range 500–750 K, QBO-associated differences in isentropic mixing are weak. However, their argument is based on comparing subtle differences in zonally averaged isentropic gradients of tracer, calculated from satellite data. Our conclusion of a strong QBO effect on isentropic mixing is supported by the results of a more detailed study, based on both numerical simulation and satellite data, to be reported in a future paper.

4. Discussion

In this paper we have analyzed the transport and mixing characteristics of stratospheric isentropic velocity fields. Our method has been to advent an artificial “test tracer” on
such isentropic surfaces and then to calculate the effective diffusivity from this tracer field.

Our results reveal clearly the height latitude variation and seasonal evolution of the stratospheric eddy transport barriers and mixing regions identified in the schematic Figure 1. It is interesting to compare our results with those of Nakamura and Ma [1997] who previously presented pictures of the height-latitude mixing of $\kappa_{\text{eff}}$, calculated from nitrous oxide ($\text{N}_2\text{O}$) fields observed by the CLAES satellite instrument. The CLAES fields are limited to the region above about 450 K (and Nakamura and Ma point out that contamination of the satellite data by Pinatubo aerosol means that there is little useful information below 600 K), but they extend to above 1900 K. The useful region partly coincides with the region considered here. In this useful region, ~600–850 K, our results and those of Nakamura and Ma are similar in showing a clear vortex/surf-zone structure in the winter hemisphere. However, there are also some substantial differences.

The first difference is that the CLAES data sometimes show a local maximum in $\kappa_{\text{eff}}$ in the tropics, particularly in summer and autumn. The second is that within the winter polar vortex, values of $\kappa_{\text{eff}}$ are high, indeed in the Southern Hemisphere higher than those in the surf zone. The third is that there are high values of $\kappa_{\text{eff}}$ at high latitudes in the summer middle stratosphere.

There are several possible explanations for these differences. One explanation is that the CLAES $\text{N}_2\text{O}$, and indeed any "real" tracer, is subject to diabatic vertical advection. If, for example, diabatic vertical advection took place on small horizontal scales, then this could have the effect of increasing the values of $\kappa_{\text{eff}}$ calculated from isentropic tracer distributions. Other possible explanations involve the limitations of the CLAES data, as discussed by Nakamura and Ma, or of the ECMWF analyses in the tropics. Also relevant is the fact that $\text{N}_2\text{O}$ is not a monotonic function of latitude and is therefore not ideal as a basis for calculating $\kappa_{\text{eff}}$; methods of accounting for the nonmonotonicity inevitably result in some anomalous values in the tropics. All these possible explanations will be discussed further in a future paper. We choose not to comment on the high-latitude differences given the lack of CLAES data coverage in this region.

Nakamura and Ma argue that the high values in the middle stratosphere summer hemisphere of $\kappa_{\text{eff}}$ calculated from observed tracers may indicate not active mixing but small-scale tracer features that have been created and "frozen in" at the time of the final warming [Hess and Holton, 1985]. In our simulations, using a diffusivity that is much larger than can be reasonably justified for the real atmosphere, these small-scale features are rapidly dissipated by diffusion. One effect of this is that high values of $\kappa_{\text{eff}}$ calculated in our simulations always represent only recent active mixing, not the effects of mixing in the distant past. Consequently, the very low values of $\kappa_{\text{eff}}$ we find in the middle stratosphere summer hemisphere indicate no recent active mixing and, together with the results of Nakamura and Ma, lend support to the Hess and Holton hypothesis that tracer features created by strong mixing at the time of the final warming can be frozen in for several months, without being dissipated by small-scale processes.

Effective diffusivity as used in this paper seems to have a number of advantages over other techniques previously used to study isentropic transport. For example, the effective diffusivity picks out the variation in tracer geometry characteristic of barriers and mixing regions wherever it occurs; the tracer does not have to be initialized in any special way to identify and quantify particular barriers. Indeed, one calculation is sufficient to analyze the transport characteristics of an entire isentropic level. Furthermore, as the flow structure evolves, for example, through the seasons, the tracer naturally adjusts to follow; no re-initialization is needed.

The method for diagnosing transport described in this paper has a number of potential future applications. One is to use minima in effective diffusivity to define transport barriers. The transport barrier defined in such a way could be used as a control surface for the calculation of fluxes of chemicals or particles, for example, in studies of the polar vortex. A second is to study the variation of the transport structure over longer timescales. In this paper we have shown that effective diffusivity can be used as an indicator of variations in transport in the tropics associated with the QBO. Such variations will be discussed further in a future paper. We have also given a first indication of interannual variability in the vortex/surf zone structure. A more comprehensive study is under way whose aim is to detect trends in Northern Hemisphere winter transport. This study will allow a more precise description of interannual variability and an assessment of connections with known sources of atmospheric variability such as the QBO. A third possible application is to use effective diffusivity as a basis for comparing transport properties of different velocity datasets and different numerical models. Differences in, for example, isolation of the tropospheric reservoir, between different datasets and different models should be clearly manifested in effective diffusivity.

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