A sudden warming in the middle atmosphere of the southern hemisphere

By D. N. AL-AJMI*, R. S. HARWOOD and T. MILES†

Department of Meteorology, University of Edinburgh

(Received 15 August 1984; revised 19 December 1984)

SUMMARY

This paper presents a case study of a wave–mean flow interaction, or minor warming, near the southern hemisphere stratopause for 15–30 July 1974. The analysis is based mainly on data from the selective chopper radiometer carried by Nimbus 5.

A deceleration of the zonal mean wind of 70 m s⁻¹ from 20 to 25 July is observed in middle latitudes at 0.5 mb, with an acceleration further south. The polar vortex is, however, not disrupted.

Eliassen–Palm cross-sections are presented for the event and a good correlation found between the E–P flux convergence and the deceleration. The channelling of the E–P fluxes from the tropopause to the convergent region is in broad agreement with the distribution of refractive index but there is a region of E–P flux divergence at high levels in polar latitudes suggesting a source of wave activity there.

The residual circulation is calculated both from the momentum budget and from an omega equation, both methods diagnosing a region of modest equatorward flow near the region of E–P flux divergence embedded in the stronger poleward flow elsewhere.

Maps of potential vorticity on the isentropic surface of θ = 1640 K are presented and the evolving pattern related to zonal mean and isobaric descriptions. The event is shown to be associated with an incursion of low latitude air into middle latitudes with a corresponding irreversible mixing of potential vorticity reminiscent of the ‘wave breaking’ in northern hemisphere events reported by McIntyre and Palmer (J. At. Ter. Ph., 46, 825), but of much smaller amplitude.

1. INTRODUCTION

(a) Motivation

There has been a considerable improvement in recent years in our understanding of stratospheric dynamics, mainly through the introduction of several interrelated concepts, notably the Eliassen–Palm flux (e.g. Edmon et al. 1980), theories of wave propagation in the meridional plane (e.g. Matsuno 1970; Palmer 1982; Karoly and Hoskins 1982), the Lagrangian average and ‘residual’ circulations (Andrews and McIntyre 1976, 1978; Dunkerton 1980) and planetary wave breaking (McIntyre 1982). These latter concepts shed considerable light on the nature of wave–mean flow interaction and the phenomenon of the sudden warming. Several observational and modelling investigations have fruitfully exploited these ideas (e.g. Dunkerton et al. 1981; Palmer 1981; Butchart et al. 1982; O’Neill and Youngblut 1982; Kanzawa 1982; McIntyre and Palmer 1983, 1984).

Studies of the sudden warming phenomenon have almost entirely been carried out for the northern hemisphere, partly for the reason that more complete synoptic data sets exist than for the southern hemisphere, and partly because the major sudden warmings of the northern hemisphere provide particularly spectacular examples of wave–mean flow interaction, unparalleled in the south. Nonetheless the southern hemisphere may be expected to show interesting variations from the northern pattern, providing a rather different regime in which the new theoretical ideas may be tested.

Several inter-hemispheric differences are already well known. In a northern hemisphere midwinter major warming, there is typically a rapid reversal of the zonal mean zonal wind from westerly to easterly through virtually the entire polar stratosphere, in association with an adiabatically induced reversal of the meridional temperature gradient. These reversals evolve over a period of a week or less; they affect the entire stratospheric and mesospheric circulation (Holton 1975), and—on at least one occasion—have been

* Present address—Dept. of Environmental Sciences, Kuwait Institute for Scientific Research, SAFAT, Kuwait
† Present address—NASA Langley Research Center, Hampton, Virginia 23665
known to extend into the lower troposphere (O'Neill and Taylor 1979). Midwinter sudden warmings of the southern hemisphere, in contrast, are confined to the middle and upper stratosphere and do not lead to a reversal of the polar night jet, despite the occurrence of local (i.e. not zonally averaged) temperature increases of similar magnitude to those reported in the north (Labitzke and van Loon 1972; Barnett 1975a).

The general inter-hemispheric difference in the behaviour of zonal mean temperature and wind is both a consequence of and a controlling factor for the inter-hemispheric difference in wave activity: noteworthy features are the smaller amplitudes in the winter southern hemisphere of quasi-stationary planetary waves and more importantly their reduced slope with height giving an almost-zero associated poleward heat flux. As the poleward heat flux is of the same sign as the vertical wave activity flux, it is clear that the southern hemisphere wave activity and vertical propagation is dominated by transient and travelling waves, in contrast to the north (van Loon and Jenne 1972; Adler 1975; Harwood 1975; Hartmann 1976a, b, 1977; Hirota 1976; Leovy and Webster 1976; van Loon 1979; Murgatroyd and O'Neill 1980; Trenberth 1980; Mechoso and Hartmann 1982; Oort 1983; Hartmann et al. 1984). For the most part these inter-hemispheric differences must ultimately depend on differences in the nature of the underlying surfaces in the two hemispheres and hence in the way waves are forced in the troposphere and propagate into the stratosphere. As consequences of these differences or at least intimately connected with them are differences in stratospheric zonal mean wind and temperature structure between hemispheres and the differences in distribution of total ozone (Dütsch 1971; Maeda and Heath 1983) which point to different Lagrangian mean circulations. A secondary but not negligible cause of inter-hemispheric differences is the variation in incident radiation brought about by the ellipticity of the earth's orbit around the sun (Barnett 1974).

With these things in mind we have studied wave growth and wave–mean flow interaction in the southern hemisphere for an unusually large warming event which occurred near 26 July 1974. This event has been discussed briefly by Barnett (1975a).

Section 2 below gives a synoptic description of the waves and the evolution of the zonal mean radiances and winds. The third section shows that the deceleration is quite well correlated with the E–P flux divergence and attempts to relate the E–P fluxes to the refractive index. Section 4 gives some estimates of the Eulerian mean and ‘residual’ circulations and in section 5 the changes in the distribution of Ertel potential vorticity on isentropic surfaces are discussed.

(b) The data

The basic data for this study are radiance measurements from the selective chopper radiometer (hereafter SCR) on the Nimbus 5 satellite. ‘Fully gridded’ data were used in which all data from one day are treated, after removal of day–night differences, as pertaining to the same map time and interpolated onto a regular (4°×10°) latitude–longitude grid. Further details of instrument performance are given by Ellis et al. (1973) and Barnett et al. (1975a), and details of the analysis method in Oxford University (1976). In addition daily analyses of the height of the 200 mb surface have been provided by the Commonwealth Bureau of Meteorology in Melbourne. The analysis method used is described in Seaman et al. (1977). Figure 1 shows the weighting function of the radiometer channels and the position of the base maps. The radiances at each grid point are converted to temperatures at 33 levels spaced at values of 0–2 in log(temperature) by a method similar to that of Rodgers (1970), except that synthetic off-diagonal values of the atmospheric covariance matrix have been used, for the reasons given by Crane (1977) (who, however, worked with Fourier coefficients) and by Miles and Chapman (1984) from whom our
values of instrumental noise are also taken. The diagonal terms of this matrix (the observational variances at each height) are based on a large sample of rocket and radiosonde data, while the off-diagonal terms are constructed by assuming a Gaussian distribution of correlation coefficients, falling to a value of +0.5 in a distance of two pressure scale heights. Four radiometer channels are used in the retrieval process, 3 peaking above the base map (B12, B34 and A2 with peaks at 1.5, 6.7 and 74 mb respectively) and one a little way below (A3, peak at 300 mb). As the lowest channels may be contaminated by cloud, the 'declouded' radiances are used in the case of the A channels. For these an upper envelope is fitted to the radiances along an orbit to determine the clear column value (Oxford University 1976). Using Melbourne analyses as base maps, heights and winds have been computed for other levels using the hydrostatic and geostrophic relationships.
2. SYNOPTIC DESCRIPTION

(a) Radiance maps

The synoptic development of the circulation of the upper stratosphere during the period of our study is shown in Figs. 2 and 3. Figure 2 shows the radiance observed on channel B12 for a period of two weeks around 26 July 1974. The situation on 17 July is typical of the period before the major wave development. There is a disturbance which can be categorized as a mixture of zonal wavenumbers 1 and 2 with the latter shown by a daily sequence to be travelling towards the east. The warm ridge near the dateline on the 17th moves east and intensifies to be centred near Argentina by 22 July. By 26 July a great change has occurred; the amplified wavenumber-2 pattern having been replaced by a wavenumber-1 pattern. The minimum temperatures are not much lower than 9 days previously but a prominent new feature is the hot area south of Africa. To illustrate further the marked geographical variation in radiance, the difference between the 26 and 21 July maps is presented in Fig. 3. A maximum temperature increase of 40 r.u. is observed in southern temperate latitudes near 0°E, whilst cooling has generally occurred from 35°S to the north pole. (1 r.u. in increment is approximately equal to 1 K.) The analysis for 30 July shows an equally dramatic variation in radiance with the virtual collapse of the wavenumber-1 hot sector leaving a predominance of somewhat smaller-scale trough-ridge features of relatively weak magnitude. The above synoptic description accords with the findings of Chapman and McGregor (1978) based on a spectral analysis designed to distinguish the travelling and stationary waves during this warming period.

Figure 2. Maps of radiance observed on channel B12 for selected days in July 1974. The contour increment is 2 radiance units (r.u.): 1 r.u. = 1 mW m⁻²str⁻¹(cm⁻¹)⁻¹. This contour increment is approximately equal to an increment in equivalent temperature of 2 K.
The B12 difference map shown in Fig. 3 also indicates the presence of zonally asymmetric wave disturbances in the ‘cooling’ region north of 35°S. The wave activity appears to stretch from 35°S to the equator and even beyond, although the amplitudes are small in the tropics. Figure 4 shows the B12 eddy radiance field on 26 July, constructed by subtracting off the zonal mean radiance at each latitude from 80°N to 80°S. This figure displays more clearly the (zonal) wave activity at the equator and in the summer hemisphere. Features with zonal peak-to-peak amplitudes of 4–5 r.u. can be seen in the summer hemisphere, well above the expected noise level in this analysis (expected noise value is 0.5 r.u.; see Al-Ajmi 1984). Near 150°W the extratropical features extend across the equator with only slightly reduced amplitude. The existence of northern hemisphere features in Fig. 4 is no proof that they are connected with southern hemisphere events of course, but the evolution seen in the full sequence of daily maps of which only part is shown in Fig. 2 suggests there is indeed a connection. Hirota (1976, 1979) has performed an analysis of SCR radiance wave activity including transient waves for 1973 and 1974,

---

Figure 3. Change in radiance observed by channel B12 from 21 to 26 July 1974. Contours are plotted at values of 0 (bold), ±1, ±2, ±3, ±4, ±5, then at multiples of 5 r.u. Positive values solid, negative values dashed.

Figure 4. The eddy radiance field observed by channel B12 on 26 July 1974 obtained by subtracting off the zonal mean radiance from the B12 radiance. Contours are plotted at multiples of 5 r.u. with the addition of those for ±0.5, ±1, ±2, ±3, ±4 r.u.
which therefore included the period of our present investigation. Westward-propagating wavenumber-1 disturbances were identified both in the mid-latitude summer hemisphere and in equatorial regions during July 1974. Barnett (1975b) has discussed cross-equatorial wave-guide effects using a similar data set but his analysis was for the quasi-stationary waves which remain after filtering out time variations shorter than about 30 days. These authors concluded that the meridional propagation of wave activity into the opposite hemisphere in the upper stratosphere is most marked at the equinoxes when the zonal mean wind is westerly. It should be noted that the waves in the cited studies were identified by filtering techniques which complicates identification with features in Fig. 4.

(b) Zonal mean radiances

The existence of dynamically induced changes in the summer hemisphere is well known from studies of the zonal mean temperatures during northern hemisphere warmings (Fritz and Soules 1970). These take the form of reductions in zonal mean temperatures at low latitudes southward of some nodal latitude, typically 45°N, simultaneously with warmings at the north pole. A similar phenomenon occurs in this southern hemisphere case but with a different meridional structure. In Fig. 5(a) we show the change in zonal mean radiance from 20 to 26 July. The radiances south of 50°S have increased by the equivalent of about 10 K. North of the nodal point at 35°S there has been a decrease of a few radiance units. (Note that in the presence of zonal variations the nodal latitude in brightness temperature may not be coincident with that in radiances due to nonlinearity of the Planck function.) Table 1 lists the two-day changes from 22 to 24 July observed in three channels. In all cases the maximum warming is at 56°S with small changes in the subtropics and cooling at low latitudes. The lower stratosphere is little changed. The main difference from a typical northern hemisphere event is that the maximum warming is in temperate latitudes rather than at the pole. The nodal point is correspondingly a few degrees equatorwards.

The explanation for equatorial coolings is complicated in an Eulerian framework by a partial cancellation between the effects of the horizontal eddy heat fluxes and of

![Figure 5](image-url)
TABLE 1. OBSERVED CHANGE IN ZONALLY AVERAGED RADIANCE (T.U.) FROM 22 TO 24 JULY

<table>
<thead>
<tr>
<th>Channel</th>
<th>Pressure of maximum weighting function (mb)</th>
<th>76</th>
<th>56</th>
<th>36</th>
<th>16</th>
<th>0</th>
</tr>
</thead>
<tbody>
<tr>
<td>B12</td>
<td>1.5</td>
<td>3.9</td>
<td>11.0</td>
<td>0.6</td>
<td>-2.2</td>
<td>-1.9</td>
</tr>
<tr>
<td>B34</td>
<td>6.7</td>
<td>2.8</td>
<td>7.0</td>
<td>0.3</td>
<td>-1.4</td>
<td>-1.2</td>
</tr>
<tr>
<td>A2</td>
<td>74</td>
<td>0.5</td>
<td>1.2</td>
<td>0.1</td>
<td>-0.2</td>
<td>-0.1</td>
</tr>
</tbody>
</table>

advection by the zonal mean vertical velocity. Dunkerton et al. (1981) have pointed out, however, that they are readily understood using Lagrangian ideas, because to the extent that the changes are rapid—and hence adiabatic—the isentropic surfaces behave like material surfaces and the total mass above a given surface must remain constant. Hence increases in pressure on an isentropic surface in high latitudes, as implied by the (Eulerian) polar warming, must be associated with reduction elsewhere. The migration of the isentropic surfaces to lower pressures is interpreted in the Eulerian framework as a cooling. The validity of these ideas has been demonstrated by Dunkerton et al. in a computer simulation of a northern hemisphere warming, supported by estimates using axisymmetric tidal theory of the cooling in the opposite hemisphere. Observational evidence for their validity in this southern hemisphere case is given in Fig. 5(b), which shows the zonal mean pressure for 20 and 26 July on the 1640 K isentropic surface which is situated a little below the 1 mb level (cf. Fig. 14). The correlations between the radiance changes in Fig. 5(a) and the pressure changes in Fig. 5(b) are obvious, while the global mean pressure is strikingly constant having changed by less than half a percent. Departure from complete constancy can be attributed to diabatic processes as well as to errors in observation, in retrieval, and in extrapolation to the poles.

(c) Zonal mean wind and deceleration

The change in the zonal mean wind associated with these temperature changes is shown in Fig. 6. Figure 6(a) shows the zonal mean westerly wind on 15 July 1974 representative of the pre-warming period. (Note that the highest level at which thermal winds can be reliably determined is 0.5 mb, about 5 km above the peak of the B12 weighting function.) The greatest wind speeds in the figure are attained at 0.5 mb in the lower mesosphere. A narrow jet core with speeds of 120 m s⁻¹ is situated near 38°S. Such intense zonal winds have been reported previously for other southern winters (e.g. Hartmann 1976a; Leovy and Webster 1976; McGregor and Chapman 1979; Hirota et al. 1983; Hartmann et al. 1984; Shiotani and Hirota 1984), but are less common in northern ‘quiet’ winter periods prior to major warmings, e.g. December. This mainly reflects the inter-hemispheric differences in seasonal wave–mean flow interaction in the extratropical stratosphere (e.g. CIRA 1972; Quiriz 1981; Hamilton 1982; Geller et al. 1983; Smith 1983; Miles and Chapman 1984). Ten days later (Fig. 6(b)) the subtropical lower mesospheric jet maximum has drifted downwards and polewards to be at 56°S and 2.5 mb.

Figure 6(c) shows a time–latitude cross-section of the zonal mean wind at 0.5 mb for sixteen days beginning on 15 July 1974. There is little change in very low latitudes, but the region of strongest wind speeds undergoes a rapid poleward shift centred on 23 July, so as to give mid-latitude decelerations and high latitude accelerations. Similar features are reported by Shiotani and Hirota (1985).

The latitude–height distribution of the zonal mean acceleration on 23 July (obtained
by differencing the wind fields of the 24th and 22nd) is shown in Fig. 6(d). The main
deceleration is at high levels and latitude 44°S, while near latitude 65°S there has been
an acceleration at high levels. This again emphasizes the contrast between this southern
hemisphere event and the typical major warming in the northern hemisphere, in that the
deceleration is at higher levels and lower latitudes than in the northern hemisphere case.
Furthermore the northern hemisphere major warmings do not normally have high latitude
accelerations. The event studied here has some similarities to the northern hemisphere
‘preconditioning event’ described by Palmer and Hsu (1983), but the height range
involved is higher for our case.

In the next section these accelerations will be compared with those implied by the
E–P flux divergence.

Figure 6. Aspects of the zonal mean wind field. (a) Zonal mean wind (m s⁻¹) for 15 July 1974. (b) As (a) but
for 25 July 1974. (c) Latitude–time section of zonal mean wind at 0·5 mb. (d) Rate of change of zonal mean
wind (m s⁻¹ day⁻¹) for 23 July 1974. Dashed contours are negative.
3. ELIASSEN–PALM FLUXES

(a) E–P flux and divergence

Several authors (e.g. Andrews and McIntyre 1976; Edmon et al. 1980) have summarized the main advantages of studying wave–mean flow interactions in terms of the Eliassen–Palm flux, namely that: (1) the E–P flux is a convenient measure of the net propagation of wave activity, being proportional to the group velocity when the latter concept applies (and only waves with a single group velocity are present at each point in the meridional plane and ray-tracing approximations to linear theory hold); (2) it is non-divergent under steady adiabatic frictionless conditions; and (3) the divergence of the flux represents the only forcing by eddies (including finite amplitudes) of the zonal mean zonal wind (at least when quasi-geostrophic scaling is appropriate).

Under quasi-geostrophic scaling the E–P flux $\mathbf{F}$ in $y, \eta$ coordinates is given by

$$\mathbf{F} = a \cos \phi \left( \frac{p}{p_0} \right) (-[u^* v^*], f[v^* \theta^*]/[\theta_\eta])$$

where $f$ is the Coriolis parameter; $\eta$ is $\ln(p_0/p)$; $p$ is pressure; $p_0 = 1000$ mb; the square bracket and asterisk denote respectively the zonal mean and departures therefrom; $a =$ radius of Earth; $u, v$ are eastwards and northwards wind components; $\phi$ is latitude; $\theta$ is potential temperature and subscript $\eta$ denotes differentiation with respect to $\eta$.

Figure 7 shows the E–P flux for 23 July. The arrows represent $\mathbf{F} \times \cos \phi$ (cf. Dunkerton et al. 1981 Eq. 2.7). Figure 7 also shows $\text{div} \mathbf{F}$ scaled to give the specific zonal force, namely $(\text{div} \mathbf{F})(p_0/\rho \cos \phi)$ (see equation after 2.5 in Palmer 1982). The data used in Fig. 7 have been filtered, only the first three waves in the zonal Fourier decomposition being retained. The contributions to the fluxes from the other wavenumbers are, however, found to be negligible (see also Holton 1975).

The forcing term shows maximum westward and eastward specific zonal force at the highest levels, respectively at mid and polar latitudes. The corresponding observed acceleration of the zonal mean wind is in Fig. 6(d). There is a very strong correlation at high levels between the forcing (Fig. 7) and the acceleration, the former being two or three times larger than the observed accelerations. This implies an upper level residual

![Figure 7](image-url)  Eliassen–Palm flux due to waves 1, 2 and 3 (arrows) and its divergence (contours) for 23 July 1974. The divergence is expressed as an equivalent acceleration in m s$^{-1}$ day$^{-1}$. Solid lines, positive divergence; dashed lines, negative divergence. See text for full definition.
circulation away from the pole at high latitudes and towards the pole in middle latitudes consistent with the displacements of the isentropic surfaces shown in Fig. 5(b), and as discussed further in section 4.

The E–P flux divergence is obviously a fairly difficult quantity to obtain from any observational data set, requiring many differentiations of the basic data. Nevertheless there are grounds for believing that Fig. 7 is at least qualitatively correct. Firstly when the E–P fluxes for each day are inspected, they exhibit a smooth temporal variation; there are quite large day-to-day changes but the pattern evolves in a consistent manner suggesting propagation of wave activity at speeds of a few km/day vertically and a few 1000 km/day meridionally. Secondly, throughout the period up to and including the 25th there is a good qualitative correlation between the sign of the E–P flux divergence and the observed change in the zonal mean wind. From the 26th to the 28th, at and after the peak of the warming, the temporal continuity of the E–P flux and especially of its divergence is reduced at great heights and high latitudes while the magnitude falls rapidly. The fluctuations appear to be associated with variations in the sign of the momentum flux which are too rapid to be fully resolved by our analysis technique. Such variations were noted also in a northern hemisphere warming by Crane (1977) and are illuminatingly

Figure 8. As Fig. 7 except time-averaged for different periods: (a) 16–19; (b) 21–24; (c) 25–28, July 1974.
commented on by Clough et al. (1985) (see also Prata 1984). A superior time analysis scheme (e.g. Salby 1982) might alleviate the problem but it is difficult to see how any observations made with a single satellite could be entirely free from these difficulties. Thirdly, there are common features with other investigations. For instance the divergence of the E–P flux in the upper stratosphere exhibits a dipole-like structure with westerly forcing poleward of the stratospheric jet core and easterly forcing on the equatorward side. Similar behaviour was observed by Hartmann et al. (1984) and Shiotani and Hirota (1985) in their analyses of southern hemisphere winters.

Rather than show the Eliassen–Palm flux for every day of this period, we have, in Fig. 8, averaged the days together according to several characteristic stages of the event as advocated by McIntyre (1982). This has the merit (in addition to reducing the amount of material to be presented) of diminishing the effects of any reversible transients. These could produce confusions of interpretation over and above the time resolution problems discussed above. Figure 8(a) shows the E–P flux in wavenumbers 1–3 and the corresponding divergence averaged for 16–19 July which can be taken to be typical of the period prior to the main interaction event. E–P flux enters the stratosphere from the troposphere in temperate latitudes. There are two main streams of activity flux, one being directed equatorwards at lower levels, the other being directed upwards and equatorwards at middle levels producing weak convergence at high levels around latitude 36°S.

The main wave forcing phase is in Fig. 8(b). The flux out of the troposphere is now much stronger and occupies a wider latitude band south of 45°S. Additionally the tendency to refract towards the equator is much reduced. In particular there is now a strong nearly vertical flux around 20 mb and it is clear that there is a great deal more wave activity reaching the middle stratosphere giving rise to the convergence in the upper stratosphere and mesosphere.

The later part of the event, including the peak temperatures, is in Fig. 8(c). Above 7 mb the E–P flux is reduced in magnitude but not greatly changed in direction, nor are the locations of the divergence and convergence greatly changed but their magnitude is reduced. Associated with this, the E–P flux from the lower stratosphere no longer appears to be feeding into the upper atmosphere to the same extent as previously. Instead the flux is beginning to revert back to the usual pattern of equatorward refraction. The vertical component of the flux at 135 mb is much reduced and the source region near 48°S has disappeared altogether, while around 40°S there is some evidence that the flux is directed back into the troposphere. This suggests that it is the disappearance of the flux through the middle stratosphere which prevents the occurrence of a wind reversal at high levels; the flux into the high stratosphere did not persist sufficiently long. Inspection of the daily values supports this conclusion, as may be seen from Fig. 9, which shows the time variation of the flux into the upper stratosphere through 22 mb. The arrows on this latitude–time cross-section are of E–P flux, plotted with the same vertical/horizontal aspect ratio as those in Fig. 8.

The strongest pulse of wave activity into the upper stratosphere takes place on 22 July when the fluxes are nearly vertical. The large fluxes persist until the 24th but are reducing by the 26th. By the 28th the fluxes are tiny and they do not recover by the 30th. As the cessation of flux into the upper stratosphere seems to be crucial to the failure of the flow to reverse, we now seek to investigate further the background to the cessation by considering the propagation characteristics of the stratosphere.

(b) Wave propagation and refractive index

The theoretical treatment of the sudden warming phenomenon has relied heavily on
linear theories in which a zonal Fourier decomposition is a convenient technique for handling aspects of behaviour which depend on the scale of the phenomena, especially wave propagation. There has been some interest therefore in studying real cases by Fourier decomposition although the amplitudes involved indicate definite limitations to whether the approach is illuminating at the peak of the warming (a manifestation of this limitation in other studies is the prevalence of large values of wave-wave interaction terms (e.g. Smith et al. 1984)).

In view of this interest we have performed an extensive Fourier analysis of the E−P flux and its divergence which we summarize in the remainder of this section. There is some indication, however, that the 200mb heights are not always adequately known for reliable results in the lower stratosphere. Trenberth (1980) writing of the 500mb monthly mean anomaly maps based on a counterpart of this same data set comments “Although wave 1 explains much of these features, the wavelength is clearly less than that of wave 1, and it seems that waves 1, 2 and 3 are all required to adequately describe the shape. For this reason, relatively small changes in shape can contribute to large changes in individual wavenumber amplitudes and phase, and owing to the large data-free areas over the Southern Oceans, the quirks of an individual [analysis] can adversely exaggerate changes in single waves from day to day”. His response to this difficulty is to refrain from looking at individual wavenumbers but to consider only wave bands comprising groups of waves. We adopt the same practice in the lower stratosphere (except Fig. 11(d) below). Fortunately the situation in the middle stratosphere and above is not so restricted. We find that the Fourier analysis of retrieved E−P fluxes and their divergences at and above 25 mb are insensitive to the assumed values of 200mb heights; they are almost unchanged by replacing 200 mb heights by their zonal mean values for instance. This is a consequence of the increase in the amplitude of height waves with height; the temperature and momentum fluxes in the upper levels are dominated by the contributions from thicknesses obtained from the more complete satellite information.

Throughout the period we find that the greatest accelerations and decelerations are produced by wavenumbers 1 and 2 with wavenumber 2 predominating from 21st to 24th. This dominance of wave 2 is, perhaps, surprising, since wave 1 has the largest temperature
amplitude by the 26th and appears to have the greatest trans-tropopause flux (not shown). Figure 10 shows the E-P flux split into wavenumbers one, two and three for 23 July. The relative magnitudes of the specific zonal force are typical of the period from 21 to 24 July inclusive. Before that the effects of waves one and two are comparable, whereas wave 1 is dominant on the 25th and afterwards, though the divergences are then reducing in magnitude.

We would obviously like to be able to explain the pulse of wave activity (Fig. 9) coming into the upper stratosphere and also the conditions responsible for its passage to high levels where the divergence is produced. Part of the explanation must involve forcing in the troposphere and lower stratosphere, which is beyond our present scope, but some insight into the stratospheric propagation in the meridional plane is provided in suitable circumstances by linear theory using for instance the ideas of ray-tracing based on the WKBJ approximation. This approximation assumes that the waves are of small amplitude so the wave equations can be linearized, and it further assumes that the meridional wavelengths of the wave are small compared to the scale of variation of the zonal mean state (Karoly and Hoskins 1982). Hence complete accord with our observations should not be expected. In these theories the eddy field is Fourier analysed into components by zonal wavenumber ($m$ say) and each component propagates separately as determined by a quantity $Q_m$, based on the zonal mean field, which will be different for each wave, where

$$Q_m^2 = (a/|u|)(\partial q/\partial \phi) - m^2/\cos^2 \phi - a^2 f^2/4N^2 H^2$$

with

$$\partial q/\partial \phi = (2(\Omega + \omega) - \omega \phi) + \omega \phi \sin \phi - (fa/NH)(p \omega)_{\eta}/p \cos \phi;$$

$\Omega =$ angular velocity of earth; $N =$ Brunt–Väisälä frequency; $H =$ pressure scale height; and $\omega = [u/a \cos \phi]$.

![Figure 10](image.png). Contributions to E-P flux and divergence for 23 July 1974 (a), from wave 1; (b), from wave 2; (c), from wave 3. The conventions are as in Figs. 7 and 8.
$Q_m$ is the 'refractive index' introduced by Matsuno (1970) from a slightly different viewpoint (but see Palmer 1982). Extra terms arise in diabatic or viscous cases, and for travelling waves $[u]$ should be Doppler shifted according to the phase speed, $c$, of the wave, i.e. replaced by $[u]-c$. Propagation is allowed if there is a sufficiently large region of positive values of $Q_m^2$ and is altogether forbidden (according to group-velocity or ray-tracing theory) in a region of negative $Q_m^2$.

Figure 11 shows the squared refractive index for wavenumber 1 (i.e. $Q_1^2$) superimposed on the same total E–P flux as in Fig. 8(a). Wave 1 has been assumed stationary in preparing this figure. $Q_2^2$ was also calculated, taking into account that wavenumber 2 is travelling throughout this event (Chapman and McGregor 1978). This was found to have a pattern very like that of $Q_1^2$, reflecting the fact that the contribution of $2\Omega - \omega_{\phi\phi}$ to $q_\phi$ is the dominant term in the expression for $Q_m^2$. However the wavenumber-2 values of $Q_2^2$ are, in general, more negative than $Q_1^2$ by typically 10 or 20 units. Figure 11 therefore gives some indication of where both waves might be expected to propagate.

![Figure 11](image-url)

Figure 11. Time-averaged E–P flux for waves 1–3 (arrows) and $Q_1^2$ (contours) where $Q_1 = \text{refractive index of wave 1}$ for (a), 16–19, (b), 21–24, (c), 25–28, July 1974; (d), as (a) but wavenumber 1 only. Dashed contours indicate negative values. The dotted line shows the positions of the zero wind line within the period.
Inspecting the refractive index, we see two main negative regions which the E–P flux vectors should have difficulty penetrating according to the theory. One of these regions is at great heights at stratopause levels in polar latitudes in the first period. The main maximum negative part at high levels moves equatorward to be at latitude 50°S by the end of the event (25–28 July, Fig. 11(c)). There is another negative region in the first period at latitude 36°S and 20 mb (Fig. 11(a)). The two negative regions are initially separated by a positive region (not present for wavenumber 2), but by Fig. 11(c), the two negative regions have merged. There is a zero wind line marked on the figure which gives infinite refractive indices. The E–P fluxes are small in the vicinity of this line.

There is a general but not perfect tendency for the E–P fluxes to avoid the negative region of refractive index squared and to be refracted towards the large values of positive squared refractive index, as would be expected from the theory. In particular the direction of flux from mid-latitude low levels in Fig. 11(b), polewards and upwards then equatorwards to mid-latitudes at stratopause levels, generally follows the ridge of higher $Q_r^2$, between the two negative extremes. There is also a measure of agreement with the diffusence of the fluxes on Fig. 11(c) into two streams, one following the positive $Q_r^2$ values back to low levels and the other travelling towards the positive $Q_r^2$ region at latitude 30°S and 1 mb. The theoretical expectation that the E–P fluxes should avoid regions of negative refractive index is not so well realized in the earlier period (Fig. 11(a)) where there is a large flux directed into the negative $Q_r^2$ (and $Q_r^2$) region at 15 mb and 36°S. This flux is in wavenumbers 2 and 3 as may be deduced by comparison with Fig. 11(d) which shows the wavenumber-1 flux alone. The E–P flux from wavenumber 1 alone follows the ridge of positive $Q_r^2$ quite closely (as it does throughout the period).

The complete interpretation of Fig. 11 is complicated by the fact that both the fluxes and the refractive index are changing in time, so that it is not easy to visualize how wave packets are behaving. Moreover even in a field of uniform $Q_1$, the wave packets should follow curved paths except in a specially chosen coordinate system (Palmer 1982). In addition the assumptions of ray theory are not strictly valid, as remarked above, both on grounds of scale and because of the presence of local sources of wave activity. It is significant for instance that we find wavenumber 2 to be strongly divergent in the lower stratosphere implying wave activity sources via wave-wave or wave–mean flow interaction, which would violate the conditions for the WKBJ approximation. While there is some doubt about how much reliance can be placed on the partitioning of fluxes by wavenumber in this region, similar patterns have been found in northern hemisphere warmings (Smith et al. 1984) so observational errors should not be too readily assumed (see also Austin and Palmer 1984).

This brief consideration of the refractive index squared appears to offer some insight into the deceleration event. In particular the general distribution of positive values seems broadly to explain the channelling of the fluxes into the convergent region. To this extent the differences from northern hemisphere events can be traced to the differences in the refractive indices (see Kanzawa (1984) for a review of previous northern hemispheric case studies). The merging of the polar mesospheric negative region with the stratospheric subtropical region is also qualitatively consistent with the observation that the E–P flux through the middle stratosphere died out before a complete reversal ensued, though tropospheric events are also relevant. The insight is only partial however; refractive index alone gives no reason for the convergent region at the middle latitude stratopause which depends on transience or dissipation, nor for that matter does it explain the location of the divergent region near the polar stratopause.
4. MERIDIONAL CIRCULATION

In this section an estimate of the mean meridional circulation at the peak of the deceleration is presented. Knowledge of this circulation is necessary to complete the understanding of the momentum and heat budgets and for the eventual explanation of features such as the differences between hemispheres of the latitude of the spring ozone maximum.

The estimates of meridional circulation presented in Fig. 12 have been calculated by an omega equation method. This uses the observation that the zonal mean winds and temperature fields are always close to thermal wind balance: the eddy fluxes of heat and momentum would destroy this balance, which must therefore be maintained by the appropriate mean meridional circulation. The stream function, \( \psi \), of the mean meridional circulation can be found by solving a second-order partial differential equation of the form

\[
L([u], [\theta]; \psi) = R_1(M) + R_2(h) + R_3(d).
\]

\( L \) is a linear second-order partial differential operator on \( \psi \), whose coefficients depend weakly on the zonal mean wind, \([u]\), and on the zonal mean potential temperature.
[θ] (mainly through its vertical derivative). L is elliptic for statically and inertially stable mean states. \( R_1, R_2 \) and \( R_3 \) are operators involving spatial derivatives and \( M, h \) and \( d \) represent respectively the convergence of eddy momentum fluxes, the convergence of eddy heat fluxes and the zonal mean diabatic heating. The formulation employed here follows that used in a circulation model by Harwood and Pyle (1975). In the southern hemisphere stratosphere and mesosphere the retrieved winds, temperatures and fluxes have been used. Elsewhere the values derived by a circulation model (Harwood and Pyle 1980) are assumed. This is thought to introduce only small uncertainties, as the fluxes in the summer hemisphere are very small and the circulation north of the equator changes by less than ten percent if the northern fluxes and zonal mean winds are set to zero in the stratosphere and mesosphere. The scheme for estimating the diabatic terms \( R_3(d) \) is as in Harwood and Pyle (1980) as far as heating by ozone absorption of UV radiation and cooling due to ozone emission of infrared radiation are concerned (including the assumed ozone amounts). \( CO_2 \) cooling in the stratosphere and lower mesosphere is calculated by a Curtis matrix method supplied by J. Haigh (see Haigh and Pyle 1982 for details).

Figure 12(a) shows the Eulerian zonal mean circulation for 23 July. The upper stratosphere is dominated by the familiar two-cell circulation comprised of a Hadley cell in low latitudes, and a Ferrel cell south of about 50°S.

The warming effects of the circulation are more readily envisaged in terms of the corresponding residual circulation \( (v_R, \omega_R) \), shown in Fig. 12(b). This was obtained from the Eulerian circulation \( (v, \omega) \) by

\[ v_R = v - \frac{p[v^*θ^*]/[θ]_η}{p} \]

and

\[ \omega_R = \frac{(v^*θ^*)/[θ]_η}{cosφ} \]

in which \( ω = Dθ/ Dt \) and subscripts \( y \) and \( η \) represent differentiation with respect to that variable.

There is basically a direct cell in the residual circulation throughout the entire upper stratosphere. This is in qualitative agreement with the requirements of the heat budget at high levels. Above about 4 mb the maximum descending motion is near 50°S, close to the latitude at which the largest rate of increase of zonal mean temperature was found. Northwards of approximately 30°S, upward velocity components are found, in qualitative agreement with the coolings. The residual circulation shown in Fig. 12(b) has a region of flow with a northward component in the vicinity of 50–70°S, 1·5 mb. This is perhaps a rather surprising feature, but as already remarked above (section 3(a)) it is also implied by the momentum budget. For comparison, the horizontal velocity component diagnosed from the momentum budget is shown in Fig. 13. This was obtained by computing \( v_R \) from

\[ -fv_R = -[u]_t + (\text{div}F)(p_0/apcosφ) \]

the right hand side being obtained from the quantities in Figs. 6(d) and 7.

Some discrepancy between Figs. 12(b) and 13 is to be expected from uncertainties in the heating rate (but see below), some from retrieval and analysis error, some from differences in the level of approximation assumed in the equations and some from finite difference errors as the two methods used grids with different resolution. In spite of this there is a qualitative agreement between the two fields in extratropical latitudes. Both have a region of northward flow at high levels and latitudes around 60°S and both have southwards flow below that in the vicinity of 3 mb. We are clearly at the limits of the
information that can be obtained from the observing and retrieval systems. However, the arguments from momentum budget and from the $\omega$ equation have a large measure of independence, so it seems fairly well established in this case that a region with a slight equatorward component of the residual circulation must have occurred at high latitudes of the upper stratosphere.

The estimates of the Eulerian mean and residual circulation presented in Figs. 12(a) and (b) both depend on an estimate of the heating rate. This can only be found if the ozone concentrations are known. In the absence of direct ozone measurements, we have used photochemical equilibrium values at stratosphere levels, and while this is a reasonable assumption (see Barnett et al. 1975b) the circulations in Figs. 12(a) and (b) are subject to some uncertainty. It is possible, however, to obtain some information on the importance of the heating terms: on account of the linearity of $L$, the mean meridional motion can be regarded as the sum of two components, one `driven' by the eddy fluxes (streamfunction, $\psi_e$, say) and one by the diabatic heating (streamfunction, $\psi_d$). These are defined by

$$L([u], [\theta]; \psi_e) = R_1(M) + R_2(h) \quad \text{and} \quad L([u], [\theta]; \psi_d) = R_3(d)$$

from which (with suitable boundary conditions) it follows that $\psi = \psi_e + \psi_d$.

We may define `the residual circulation driven by the eddies' (streamfunction $\psi_{re}$, say) by subtracting the motion driven by the heating from the residual circulation, thus $\psi_{re} = \psi_e - \psi_d$. It is readily seen from the linearity of $L$ that $\psi_{re}$ does not depend on our estimate of the heating rate. The circulations corresponding to $\psi_{re}$ and $\psi_d$ are shown in Figs. 12(c) and (d) respectively. From Figs. 12(c) and (d) it can be seen that the equatorward components of the flow near 50$^\circ$S are attributable to the eddy behaviour, the part of the motion driven by the heating having slight poleward components in this region (and likely to for any reasonable estimates of ozone concentration). It is also apparent from Fig. 12(c) that the effect of the eddies in the southern hemisphere drives a circulation which extends well into the opposite hemisphere. This is consistent, of course, with the zonal mean coolings noted (e.g. in Fig. 5) north of about 40$^\circ$S and with theoretical expectations (Dunkerton et al. 1981, Fig. 10).
It is noteworthy that the residual circulation produced by the eddies makes a much larger contribution to total residual circulation on this day (and throughout the period 15–30 July) than the portion produced by diabatic heating. This throws considerable doubt on the modelling assumption adopted by Rogers and Pyle (1984) that the part driven by the heating can be treated as the entire residual circulation, though their assumption may of course be reasonable when averages over longer time periods are required.

5. Analysis on isentropic surfaces

In this section we present some features of the interaction analysed on potential temperature surfaces, in particular we present maps of potential vorticity to seek signs of irreversible mixing of the type suggested by McIntyre and Palmer (1983, 1984). The surface chosen for study is \( \theta = 1640 \text{K} \) which is near the 1·5 mb level. Geostrophic winds and hence an estimate, \( Z_m \), of the Ertel potential vorticity appropriate for large Richardson numbers were obtained from the Montgomery potential \( \psi_m \), where \( \psi_m = c_p T + gz \), \( g \) being the gravitational acceleration and

\[
Z_m = -\left(\frac{g}{\theta}\right)\left(\frac{\partial \theta}{\partial \rho}\right)\left[\left\{\frac{1}{f}\right\} + \left(u/a\right)\frac{\partial f}{\partial \phi}\right] + f. \]

Figure 14 shows the winds and pressure field on the \( \theta = 1640 \text{K} \) isentropic surface for two days, one before the main interaction event and one at the peak of the event. As noise in the Montgomery field can produce large errors in the low latitude geostrophic winds, the velocities equatorwards of 8°S have been set to zero. This procedure may be unnecessarily pessimistic, however, as a comparison with the available rocket soundings in the period of our study (see Table 2) shows good agreement not only in middle and high latitudes but also at Ascension Island, 8°S.

There are two pressure minima a little under 1·1 mb on the chart for the 15th and these correspond, as they must, to two BI2 radiance minima for this day, which can be seen slightly displaced two days later in Fig. 2(a). Likewise the pressure maxima on Fig. 14(a) correspond to hot regions on the radiance maps. The winds are of the order of 100 m s\(^{-1}\) in mid-latitudes, taking roughly two days to circle the pole at 60°S. The winds are strikingly zonal on the 15th polewards of approximately 40°S, and detailed examination confirms that this figure is consistent with the wind cross-sections shown in Fig. 6(a) above.

The complete sequence of charts shows a small anticyclone developing on the 22nd near 30°W, which is distinct from the small anticyclonic circulation to be seen on the 15th at 25°S 10°E. By the 26th (Fig. 14(b)) this new anticyclone has enlarged considerably

<table>
<thead>
<tr>
<th>Station</th>
<th>Date July 1974</th>
<th>Pressure or height</th>
<th>Rocket wind</th>
<th>Satellite geostrophic wind</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Direction °</td>
<td>Speed (m s(^{-1}))</td>
</tr>
<tr>
<td>Moledezhnaia (67°S 46°E)</td>
<td>17</td>
<td>137 Pa</td>
<td>273</td>
<td>76</td>
</tr>
<tr>
<td>Ascension Island (8°S 74°W)</td>
<td>17</td>
<td>155 Pa</td>
<td>267</td>
<td>118</td>
</tr>
<tr>
<td></td>
<td>23</td>
<td>150 Pa</td>
<td>086</td>
<td>40</td>
</tr>
<tr>
<td></td>
<td>26</td>
<td>150 Pa</td>
<td>088</td>
<td>30</td>
</tr>
<tr>
<td>Mar Chiquita (38°S 57°W)</td>
<td>17</td>
<td>44 km</td>
<td>278</td>
<td>116</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>094</td>
<td>38</td>
</tr>
</tbody>
</table>

TABLE 2. Comparison of rocket- and satellite-derived winds at 1650 K potential temperature surface
Figure 14. Winds and pressure field on the isentropic surface, $\theta = 1640$ K. (a) 15, (b) 26, July 1974. The scaling of the wind vectors is shown by the arrow to the lower right of the figure which corresponds to 20000 km day$^{-1}$, i.e. approximately 200 m s$^{-1}$. Pressure contours are labelled in pascals.
and is centred near 35°S 10°E, dominating the flow in mid and low latitudes from 60°W, eastwards to 110°E. The deceleration of the zonal mean wind seen in Fig. 6 is clearly associated with this feature. The topography of the surface is also much altered, with a single pressure maximum slightly over 1.9 mb at 65°S 15°E, corresponding to the channel B12 radiance maximum in Fig. 2. The winds in the middle of this high pressure are stronger than they were on the 15th, attaining velocities of 160 m s⁻¹. In high latitudes the polar vortex is essentially intact though it is somewhat elongated.

Figure 15 contains maps of Ertel potential vorticity before the event ((a), 15 July), near to the maximum deceleration ((b), 24 July) and after the main deceleration ((c), 28 July). In preparing Fig. 15 the potential vorticity was calculated on a latitude–longitude grid extending to 76°S. This was subjected to a weak smoothing operation (Wallington 1962) prior to interpolation onto the polar stereographic map for plotting, at which stage the polar values were obtained by extrapolation. Equatorwards of 8°S the relative vorticity (not potential) has been set to zero because of possible contamination of low latitude geostrophic winds by noise in the Montgomery potential, though as remarked above this procedure may be unnecessarily stringent.

The lowest (most negative) and highest values of potential vorticity vary little in this two-week period (though strict conservation is not expected over this length of time at these heights) but there has been a great reduction in the areas occupied by the low potential vorticity and a marked poleward shift in the region of large gradients between the 15th and the 28th.

The redistribution can clearly be seen taking place in Fig. 16, which shows a latitude–
time cross-section of the zonal mean potential vorticity and the corresponding eddy flux of potential vorticity. On the 15th the zonal mean potential vorticity has large gradients near 70°S and from 25° to 45°S, with a small or slightly reversed gradient between. By the 24th the gradient is fairly uniform with latitude although the zonal mean picture is really quite misleading on this day as the gradients are still strong at any given longitude, but the eccentricity and elongation of the potential vorticity contours produce a smearing out of the gradients when zonally averaged (cf. Fig. 15(b) for the 24th and Fig. 17 below).

In view of (a), the definition of refractive index (see section 3 above) which involves the poleward gradient of the zonal mean quasi-geostrophic potential vorticity and (b), the close analogy between, on the one hand, quasi-geostrophic potential vorticity on pressure surfaces and, on the other hand, Ertel potential vorticity on isentropic surfaces (Charney and Stern 1962), it is to be expected that the regions of positive squared refractive index in Fig. 11 should be related to large gradients in Figs. 15(a) and 16(a). This is indeed the case: the positive values in Fig. 11(a) near 1.5 mb lying between 28°S and 48°S are clearly related to the strong gradients in that region seen on Fig. 15(a). The subsequent collapse of this positive region (Figs. 11(b) and (c)) is related to the redistribution of the potential vorticity which produces reduced gradients in Fig. 16(a).

A comparison of the time change of zonal mean wind (Fig. 6(c)) and the fluxes in Fig. 16(b) gives qualitative confirmation of the expectation from quasi-geostrophic theory that the deceleration of the zonal mean wind should be correlated with the southward flux of potential vorticity (being proportional to the convergence of E–P flux). The poleward flux of potential vorticity can conveniently be thought of in the southern hemisphere as an equatorward flux of negative potential vorticity for analogy with the northern hemisphere. Both the latitude and timing of the maximum eddy flux are well correlated with the zonal mean deceleration. The fluxes are also well correlated with the change in the gradient of zonal mean potential vorticity, even though the effect of the mean meridional wind, [v][Z], is comparable to the eddy fluxes, as may be deduced by noting that the residual velocities in Figs. 12 and 13 are a few metres per second.
Figure 15. Ertel potential vorticity on the isentropic surface $\theta = 1640\, K$. (a) 15, (b) 24, (c) 28, July 1974. The contours plotted are $(Z/g) \times 10^{-6} \text{Pa}^{-1}\text{s}^{-1}$). See text for definition.
Figure 15. (continued)

Figure 16. (a) Zonal mean potential vorticity on the isentropic surface, $\theta = 1640\, K$. Contours are of $(Z/g) \left(10^{-4}\, \text{Pa}\, \text{s}^{-1}\right)$. (b) Zonal mean northward flux of potential vorticity on the same isentropic surface. Contours are of $(\cos\phi[v \cdot Z]/g) \left(10^{-3}\, \text{Pa}\, \text{m}\, \text{s}^{-1}\right)$. $\phi$ is latitude.
Figure 17. As Fig. 15 except only contour values $-6$, $-8$, ..., $-16$ are shown. (a) 15, (b) 22, (c) 24, (d) 26, (e) 28, (f) 30, July 1974.
Figure 17. (continued)
The potential vorticity distributions in Fig. 15 show the main vortex to be distorted but apparently remaining intact. We find, however, that mixing does take place on the equatorward edge of the main vortex through parcels with low (large negative) vorticity air being separated from the main vortex. This process is active in particular in producing the areas of high potential vorticity near 30°E 50°S on the 28th and the associated plateau in the contours.

The mixing is more discernible in a sequence of maps with only a limited range of contours plotted but with reduced contour interval as in Fig. 17, on which only contours at intervals of two units from 6 to 16 units appear (cf. Fig. 15 where spacing was 5 units).

On the 15th (Fig. 17(a)) (included for comparison with 15(a)) the contours are rather smooth and at nearly all latitudes the potential vorticity decreases monotonically towards the pole (within the range of values plotted). By the 22nd the vortex appears less smooth. The intervening maps (not shown) provide some evidence in low latitudes of portions of the vortex becoming detached and interchanged with air from the equatorial region, but the problem of obtaining reliable estimates geostrophically in low latitudes leaves room for doubt. The remarks of McIntyre and Palmer (1983) that the potential vorticity maps from satellite data inevitably are like looking at reality 'through knobbly glass' apply here with equal force. However, the development between the 22nd and the 30th of the pattern in the sector from 0° to 90°E in the vicinity of 50°S shows convincing continuity from day-to-day as can be surmised from the sequence of maps at two-day intervals presented here. Hence there can be little doubt that the charts show 'strips' of air of subtropical or tropical origin being injected polewards and eastwards into the main vortex. By the end of the sequence the subtropical air is losing its identity and 'merging' with the higher latitude background values. With the type of data at our disposal it is impossible to see whether the potential vorticity has altered through frictional and/or radiative processes to attain values comparable with the new environment, or whether it still exists with unaltered potential vorticity as a very narrow strip, too fine to be resolved. The incursion of the air from subtropical to high latitudes is broadly consistent with the winds around the anticyclone centred at 35°S 10°E on 26 July. It would be interesting to compute trajectories to determine whether the air has passed round the anticyclone, originating from the easterly flow on its northern flank, or whether it originated in the westerlies of the low latitudes of the western hemisphere. Unfortunately the present analysis scheme is probably not capable of reliably determining such a comparatively subtle issue.

6. Summary and conclusions

This case study of a southern hemisphere warming has shown several differences from a typical northern hemisphere case. Prominent amongst these is the fact that the deceleration took place at higher altitudes and lower latitudes than in northern hemisphere events. The magnitude of the reduction of zonal mean westerlies was comparable with those found in northern hemispheric events and yet the wind failed to reverse, the initial wind strengths being considerably larger than in the northern case. The failure of the wind to reverse may also be attributed more fundamentally to the reduction of the E–P fluxes passing through the middle stratosphere to the region of convergence.

The E–P fluxes and the associated divergence and convergence are mainly in wavenumbers one and two above the middle stratosphere with only small contributions from wavenumbers three and higher. The refractive indices for these waves have a large measure of consistency with the behaviour of the fluxes, especially in the case of wavenumber one and in spite of the large amplitudes which would cast doubt on the
applicability of linear theory. In particular the channelling of the fluxes shows a good qualitative agreement with high values of the refractive index squared, especially in the case of wavenumber one. Moreover the cutting off of the fluxes through the middle stratosphere towards the end of the period occurs in association with the replacement of the real by imaginary values of refractive index in that region. However, events in the troposphere are also important in the decay of the fluxes.

In common with several other studies we find a region of E–P flux divergence at high latitudes. The reason for this divergence is not immediately clear but it does suggest a source of wave activity, as has been discussed by Hartmann (1983), who attributes it to a barotropic instability. We do find limited evidence for this idea in weakly reversed gradients of potential vorticity in the zonal mean and more strongly reversed gradients locally but these appear to be the result of the wave motions rather than their cause.

The residual circulation is in a sense to offset the effects of both the E–P flux divergence and convergence (as expected from Eliassen's (1951) theory of forced meridional circulations) and therefore has a small component towards the equator in high latitudes of the southern hemisphere at great heights but towards the south pole elsewhere. This former regime is considerably different from the residual circulation published in most other studies, but the same sense of this circulation is obtained by two essentially independent calculations and so appears to be firmly diagnosed. The vertical velocities of the residual circulation and especially the part attributable to the eddy forcing is consistent with the zonal mean heating in mid-latitudes and cooling to the north.

Analysing the warming on isentropic surfaces reveals in more detail many of the processes relevant in the zonal mean description, namely the structure of the systems producing the potential vorticity fluxes, which are intimately connected both to the E–P flux convergence and to the changes in the zonal mean potential vorticity gradient, which is in turn of great importance in the refractive index. An incursion of air from low latitudes into middle latitudes leading to an apparent mixing of the potential vorticity can be detected, reminiscent of the processes described by McIntyre and Palmer (1983 and 1984), but it is more localized than in the northern hemisphere major warming events and does not lead to a total disruption of the polar vortex.

One strong impression which arises from analysing the maps is of the obscuring effect of dealing in terms of the zonal mean: at each longitude there are strong meridional gradients of potential vorticity, yet the elongation of the vortex smears this gradient over a wide band of latitudes with correspondingly reduced magnitude when considered as a zonal mean. Moreover the need to explain the zonal mean deceleration is seen on hemispheric maps to translate into a question of what is responsible for the growth of a rather localized anticyclone in subtropical latitudes. It is not that the E–P cross-section is in error in its information concerning the zonal mean deceleration, it is rather that it is not particularly illuminating when the phenomenon under investigation is ill portrayed as a small perturbation from a zonally symmetric state. This is of course widely recognized. However, notwithstanding some potentially useful suggestions by McIntyre (1982) and by Palmer and Hsu (1983) concerning ways of dealing with extremely asymmetrical states, totally compelling alternative theoretical concepts are not yet apparent.

ACKNOWLEDGMENTS

We are indebted to Ms B. Naujokat of the Free University of Berlin for the rocketsonde data and to Dr M. E. McIntyre for useful comments on details of an earlier version of the manuscript.
This work was funded by the Natural Environment Research Council. D. N. Al-Ajmi wishes to thank the Kuwait Institute for Scientific Research for financial support.

REFERENCES


Barnett, J. J. 1974 The mean meridional temperature behaviour of the stratosphere from November 1970 to November 1971 derived from measurements by the Selective Chopper Radiometer on Nimbus IV. *ibid.*, 100, 505-530


CIRA 1972 *COSPAR International Reference Atmosphere*, Akademie Verlag, Berlin


Eliassen, A. 1951 Slow thermally or frictionally forced meridional circulations in a circular vortex. *Astrophysica Norvegica*, 5, 19-60


Fritz, S. and Soules, S. D. 1970 Large scale temperature changes in the stratosphere observed from Nimbus III. *J. Atmos. Sci.*, 27, 1091-1097


Hartmann, D. L. 1976a The structure of the stratosphere in the southern hemisphere during late winter 1973 as observed by satellite. ibid., 33, 1141–1154

1976b The dynamical climatology of the stratosphere in the southern hemisphere during late winter 1973. ibid., 33, 1789–1802


1983 Barotropic instability of the polar night jet stream. J. Atmos. Sci., 40, 817–835


Harwood, R. S. and Pyle, J. A. 1975 A two-dimensional mean circulation model for the atmosphere below 80 km. ibid., 101, 723–747

1980 The dynamical behaviour of a two-dimensional model of the stratosphere. ibid., 106, 395–420

Hirota, I. 1976 Seasonal variation of planetary waves in the stratosphere observed by the Nimbus 5 SCR. ibid., 102, 757–770

1979 Kelvin waves in the equatorial middle atmosphere observed by the Nimbus 5 SCR. J. Atmos. Sci., 36, 217–222


Palmer, T. N. Oxford University 1976 ‘The Nimbus V Selective Chopper radiometer’. Atmospheric Physics Memorandum 76.1, Clarendon Laboratory


Prata, A. J. 1984 The 4-day wave. *ibid.*, **41**, 150–155


