The Quasi-Two-Day Wave Event of January 1984 and Its Impact on the Mean Mesospheric Circulation

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ABSTRACT

Studies of the quasi-two-day wave show that it is a summertime phenomenon. In the summer of 1983–84 at Adelaide (35°S, 138°E), the main phase of the wave appeared as a pulse in mid-January which lasted about seven cycles (14 days). Coincident with the onset of the pulse a temporary but substantial change occurred in the prevailing circulation throughout a deep layer of the upper mesosphere; a perturbation of more than 10 m s^{-1} occurred in the northward flow, whereas the change in the zonal flow (about 30 m s^{-1} westward) actually caused a reversal of the prevailing eastward circulation above 84 km.

It is suggested that these changes in the prevailing circulation were a response to the wave pulse. A simple calculation is performed to estimate the anticipated response to the observed wave event; under plausible assumptions about the magnitude of mean and eddy dissipation processes, predicted circulation changes agree reasonably well with those observed.

It is concluded that such events have a substantial, if temporary, impact on the prevailing circulation in the upper mesosphere and may be important in the transport of atmospheric constituents at these heights during summer.

1. Introduction

A global-scale oscillation with about a 2-day period is now recognized to be an ubiquitous feature of the summer upper stratosphere and mesosphere. It has been observed with ground-based radars in studies of mesospheric winds (Kingsley et al., 1978; Craig and Elford, 1981; Salby and Roper, 1980; Stening et al., 1978; Glass et al., 1975) and in satellite observations of stratospheric temperatures (Rodgers and Prata, 1981). Simultaneous observations made at more than one longitude show that the wave is westward traveling with a zonal wavenumber of 3 (Muller and Nelson, 1978; Craig et al., 1980; Rodgers and Prata, 1981). The observations also show that the wave is strongly asymmetric, being confined mainly to the summer hemisphere and reaching its strongest amplitudes just after the summer solstice, about July/August in the Northern Hemisphere and in January in the Southern Hemisphere. The wind oscillations typically maximize in the meridional component at low- to midlatitudes (Vincent, 1984) with maximum amplitudes usually attained at heights between 80 and 90 km. Radar wind measurements reported by Craig et al. (1983) also indicate that the wave extends from the summer into the winter hemisphere. They found that quasi-two-day mesospheric wind oscillations observed in July and August 1981 at 53°N in the Northern Hemisphere were observed simultaneously at 19°S but not at 35°S. The satellite measurements show that the maximum temperature oscillations are reached in the mesosphere at summer latitudes near 20°, with the wave structure extending from near 50° in the summer hemisphere to about 20° latitude in the winter hemisphere (Rodgers and Prata, 1981). Both satellite and radar studies show that the wave is almost evanescent in the vertical although some phase tilt with height, consistent with upward propagation, is found in high-resolution radar measurements. Vertical wavelengths of over 100 km are usual although shorter wavelengths are inferred at low latitudes (Craig and Elford, 1981; Craig et al., 1980; Craig et al., 1983).

The features described here are common to the observations made in both the Northern and Southern Hemispheres. It should be recognized, however, that there are some hemispheric differences. Most importantly, the wave amplitudes are usually much larger in the Southern Hemisphere. Observations made at Adelaide (35°S) give a typical mean maximum amplitude in the meridional wind component of about 35 m s^{-1}

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(Vincent, 1984) but short-term amplitudes of between 50 and 100 m s^{-1} are not uncommon (Craig and Elford, 1981; Craig et al., 1980). In contrast, maximum amplitudes near 20 m s^{-1} are typical of Northern Hemisphere observations (e.g., Vincent, 1984). Rodgers and Prata (1981) also found that the temperature amplitudes were twice as large in the Southern Hemisphere summer as in the Northern Hemisphere summer. The wave periods also appear to differ somewhat between hemispheres. Southern Hemisphere mesospheric wind observations all give periods close to 48 h (Craig and Elford, 1981; Craig et al., 1980), whereas periods near 52 h are found in the Northern Hemisphere (Kingsley et al., 1978; Salby and Roper, 1980; Stening et al., 1978; Glass et al., 1975). It is possible that these differences are due to hemispheric differences in the wave forcing and/or propagation conditions in the middle atmosphere. Salby (1981, 1984) suggests that the two-day wave is a manifestation of the (3,3) Rossby-gravity normal mode forced in the lower atmosphere. Plumb (1983) and Pfister (1985) have examined the possibility that it is caused by a baroclinic instability of the summer easterly wind jet.

In this paper we report new observations of the 2-day wave made at Adelaide in the Southern Hemisphere summer of 1983/84. These observations were made as part of a program of continuous monitoring of the dynamics of the mesosphere and so differ from previous measurements made at Adelaide which were of short duration, never exceeding a few weeks in length. As in previous years, we observed a large-amplitude 2-day wave event in January 1984 and the continuity of the observations has enabled an investigation to be conducted of the changes in the prevailing circulation during this event. It is argued that a significant wave-mean flow interaction occurred.

2. Data

The wind data were obtained using the Buckland Park 2-MHz radar located 40 km north of Adelaide (35°S, 138°E). The radar has measured the winds continuously since November 1983, using the spaced antenna technique (e.g., Vincent, 1984) in the height range 60–100 km during the day and 78–100 km range at night, with a height resolution of 2 km and time resolution of a few minutes. For the results discussed here, hourly mean values were used.

To illustrate the main features of the 2-day wave in 1983–84, the data were subjected to a band-pass filter with cutoff periods of 30 and 80 h. Figure 1 shows the resulting filtered wind for the zonal (EW) and meridional (NS) components in the 82–98 km height range for a 54-day interval commencing 1 January. It is apparent that the wave is transient in nature with the largest event taking the form of a pulse or burst starting in mid-January and lasting for about 7 cycles. A higher-resolution plot for the period 20–30 January is shown in Fig. 2. As is usual at Adelaide, the NS amplitudes, which reach peak values in excess of 50 m s^{-1}, are appreciably larger than those for the EW component. It is also noticeable that there is an appreciable phase

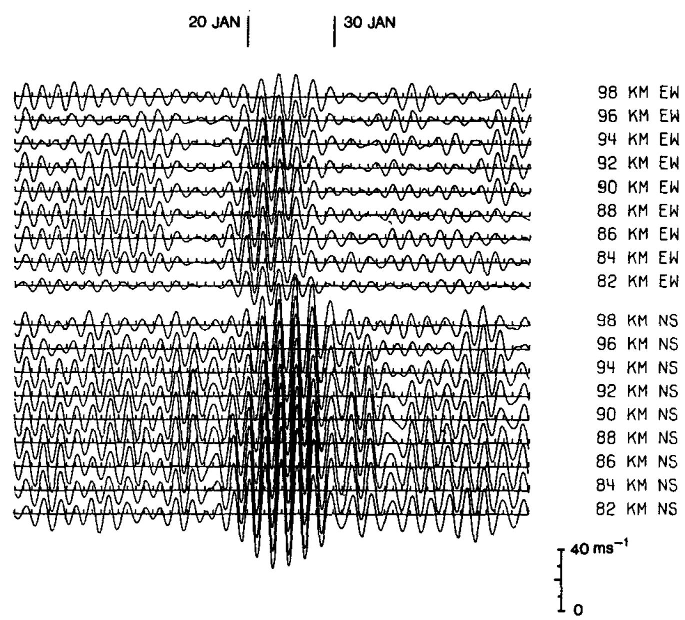

FIG. 1. Eastward (top) and northward (bottom) components of the winds observed at Adelaide between 82 and 98 km during January and early February 1984. Data have been band-pass filtered to retain only periods between 30 and 80 h. Scale at lower right. Tick marks on axes denote daily intervals.

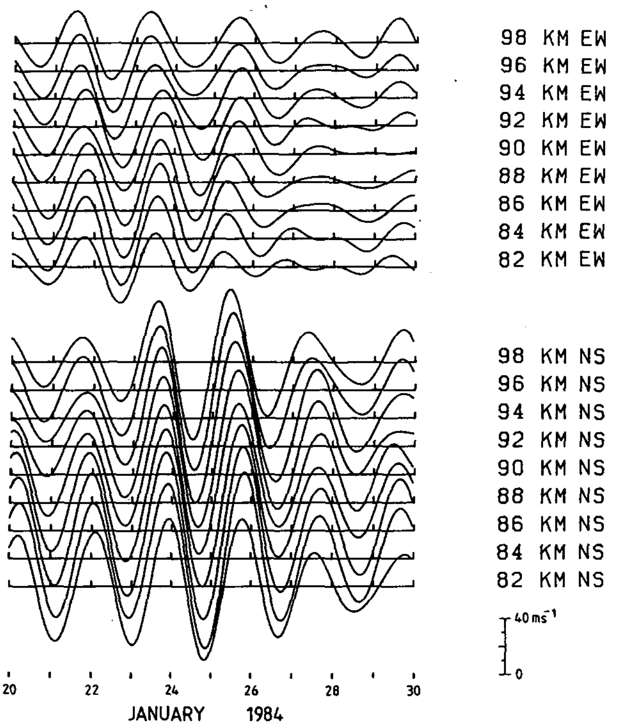

FIG. 2. Expanded plot of Fig. 1 for the period 20–30 January 1984.

change with height of the NS oscillations, the sense of the tilt indicating upward propagation with a vertical wavelength in excess of 100 km. The phase tilt is smaller in the EW component so that the phase difference between the EW and NS oscillations changes with increasing height; nevertheless, they are in approximate phase quadrature at the lower levels. Again, these results are in accord with previous observations of the 2-day wave at southern midlatitudes (Craig et al., 1980).

The prevailing winds which were obtained in January 1984 are typical of this summer period, with easterlies reaching maximum values of about 70–80 m s^{-1} in the lower mesosphere and westerlies of about 25 m s^{-1} at heights near 100 km (Fig. 3). The meridional circulation was basically equatorward, with the maximum flow occurring at 85 km, the height where the zonal flow was about zero. A detailed examination of the mean winds showed that during January the zonal circulation was steadily evolving with time as shown in Fig. 4a for the 70–98 km height interval. The maximum easterly velocities were attained just after the solstice and thereafter became more westerly while at the uppermost levels the reverse occurred. However, for a period of several days near the time of the large burst in two-day wave activity in mid-January this steady change in the zonal flow was dramatically reversed, especially at heights near 86 km although the effects were evident at heights 10 km above and below this level. Corresponding but smaller changes were also seen in the meridional circulation (Fig. 4b). The band-pass filtered NS wind component and the low-pass filtered EW and NS components (with the trends removed) observed at 86 km are plotted in Fig. 5 to bring out more clearly the apparent association between the two-day event and the changes in the prevailing winds.

3. Impact of the wave event on the zonal mean flow

The changes in low-passed eastward and northward wind evident around 20 January (Figs. 4 and 5) appear to be significant features in the record. The fact that these changes are approximately coincident with the onset of the wave pulse suggests a relation between the two; here we investigate the possibility that the low-pass wind changes are a manifestation of the impact of the wave event on the background circulation.

Our ability to pursue a thorough analysis is limited by the nonavailability of observations at other sites. However, much can be elucidated if we make the assumption that the Adelaide observations are typical of all similar longitudes in the sense that the time averages and departures therefrom in these observations may be equated to the zonal average and departures therefrom around 35°S latitude. While we are not aware of any other observations of the 1983–84 event to support this contention, analyses of observations in previous

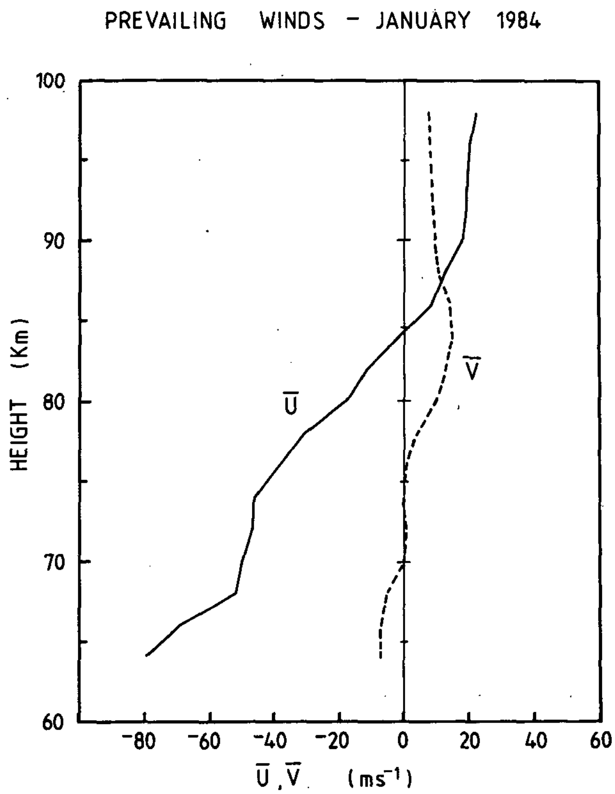

FIG. 3. Monthly mean eastward (solid) and northward (dashed) wind components observed at Adelaide for January 1984.

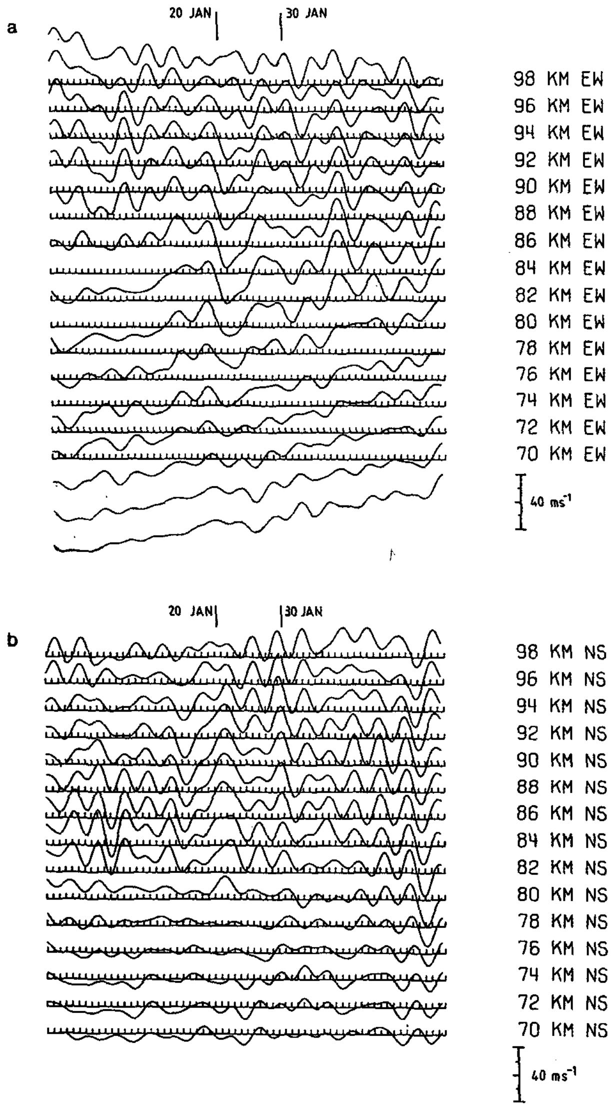

FIG. 4. (a) Eastward and (b) northward components of the winds observed at Adelaide between 82 and 98 km during January and early February 1984. Data have been low-pass filtered to retain only periods longer than 80 h. Scale at lower right. Tick marks on axes denote daily intervals.

years from multiple sites (Craig et al., 1980) and from satellite measurements (Rodgers and Prata, 1981) have identified the quasi-two-day wave as a zonally propagating wave of zonal wavenumber 3. The present observations appear to be characteristic of those made in previous years. Further, it appears unlikely that the pulse seen at Adelaide is a local view of a spatially localized disturbance being advected past the observing site, since it is difficult to envisage such a disturbance maintaining its vertical coherence in the strong mean shear evident in Fig. 3. Therefore, we believe this hypothesis to be consistent with our understanding of the phenomenon.

We take a simple Gaussian description of the wave pulse by writing

$$v' = \mathrm{Re}\, V e^{ik(x-ct)} \exp\left(\frac{-t^2}{2T^2}\right) \tag{1}$$

where $t = 0$ on 26 January, $T = 7.5$ day and V is in the range 50–60 m s^{-1} between 82 and 98 km, v is the northward velocity, the notation $(\ )'$ denotes departure from a zonal average and c is the zonal phase velocity of the wave (about -60 m s^{-1} at 35°S for a westward wavenumber 3 wave of a 2-day period). If η is the northward parcel displacement in this wave field, then linear theory tells us that

$$\left(\frac{\partial}{\partial t} + \bar{u}\frac{\partial}{\partial x}\right)\eta = v' \tag{2}$$

where $\bar{u}$ is the mean zonal wind. Using this relation, (1) gives

$$\eta = \mathrm{Re}\, Y e^{ik(x-ct)} \exp\left(\frac{-t^2}{2T^2}\right) \tag{3}$$

where we estimate that $|Y|$ decreases from about 3000 km at 82 km altitude to about 1200 km at 98 km. These estimates are based on $\bar{u}$ values given by the observed monthly mean shown in Fig. 3; the 10-day averaged winds over the interval 20–30 January are, in fact, little different from these. Note that $|Y|$ may become even larger below 82 km; there is, in fact, a critical line ($\bar{u} = c$) at around 67 km (see Fig. 3). Such enormous excursions of material parcels (up to 60 deg latitude peak-to-peak) serve to illustrate the amplitude of the wave at these levels during southern summer. Furthermore, it also leads us to question the validity of the linear assumption in (2). We shall return to this point later.

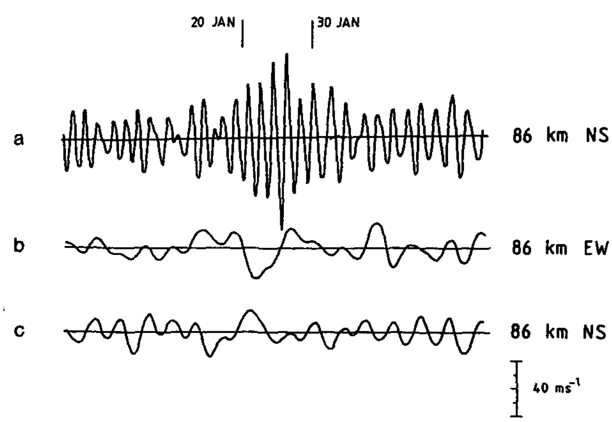

FIG. 5. The wave pulse at 86 km. Band-pass northward velocity and low-pass eastward and northward velocities (with mean and trend removed) at 86 km for the same interval of time shown in Figs. 1 and 4. Scale at lower right.

The quasi-geostrophic momentum budget for the zonal mean flow may be written

$$\frac{\partial}{\partial t}(\delta\bar{u}) - f\delta\bar{v} = -\frac{\partial}{\partial y}(\overline{u'v'}) + \bar{X} \tag{4}$$

where $\delta\bar{u}$ and $\delta\bar{v}$ are the changes in mean eastward and northward wind, respectively, during the event (defined such that $\delta\bar{u} = \delta\bar{v} = 0$ prior to the wave event), $f = 2\Omega$ sin (latitude) is the Coriolis parameter and X is a body force (per unit mass) representing the effects of gravity wave breaking and turbulence. Rather than using a sophisticated parameterization scheme for the latter, we prefer to follow an earlier practice of representing this force crudely as a Rayleigh friction, namely,

$$\bar{X} = -\lambda\delta\bar{u}. \tag{5}$$

The value of λ will be addressed here. Since $\bar{v} \approx 10$ m s^{-1} during the event (Fig. 5), the Coriolis term in (4) would, alone, generate zonal accelerations of about 70 m s^{-1}/day, far in excess of what is observed. Therefore, this acceleration must be balanced by either or both of the terms on the right-hand side of (4). In fact, from the observed band-passed winds of Fig. 1, we note that $|u'| \leqslant 20$ m s^{-1}, $|v'| \leqslant 60$ m s^{-1} and, furthermore, that the two are approximately in quadrature over the lower portion of the height range. In fact, the maximum $|\overline{u'v'}|$ (at the peak of the wave pulse) is about 800 m^2 s^{-2} at 90 km, falling off weakly above, and rapidly below this level. Taking a value 5000 km as a latitudinal length scale [again we resort here to observations of previous events, particularly the satellite observations of Rodgers and Prata (1981)], we estimate that $\partial(\overline{u'v'})/\partial y$ is less than 20 m s^{-1}/day (and much less than this below 90 km). Therefore, it seems that the momentum balance as described by (4) during the event is primarily a balance between Coriolis and frictional effects; from the low-pass winds shown in Fig. 2, (4) demands a Rayleigh friction coefficient $\lambda \approx (0.5\text{ day})^{-1}$, a value rather greater than most theoretical estimates (e.g., Miyahara 1985).

Of course, the weak contribution of the eddy momentum flux in (4) does not imply there is no significant role for the wave in the momentum budget, since the mean circulation $\delta\bar{v}$ could be driven by the eddy heat fluxes. Indeed, if the changes in the low-pass winds that occurred at the time of the wave event were, in fact, caused by that event this must have been so. We could, in principle, calculate this effect through analysis of the mean heat budget. However, the impact of the wave on the zonal mean state is more easily assessed using the transformed Eulerian-mean form of the quasi-geostrophic momentum budget (e.g., Edmon et al., 1980)

$$\frac{\partial}{\partial t}(\delta\bar{u}) - f\delta\bar{v}_* = \overline{v'q'} + \bar{X} \tag{6}$$

where $\delta\bar{v}_*$ is the change in the "residual" northward flow and q is the quasi-geostrophic potential vorticity. For a deep system such as the one we are considering here (cf. Figs. 1 and 4), the Coriolis term is usually small. In general, the primary role of the residual circulation is to effect a vertical redistribution of the response to localized forcing (e.g., McIntyre 1980). Continuity demands that the Coriolis term in (6) vanish if we multiply by pressure and take a vertical integral over a sufficiently deep layer, i.e., over a layer significantly deeper than fL/N, where L is a latitudinal length scale and N the buoyancy frequency. For $L \approx 5000$ km, $fL/N \approx 25$ km. Data limitations preclude our taking data from such a deep layer; however, it seems reasonable to write

$$\frac{\partial}{\partial t}(\delta\bar{u}) \approx \overline{v'q'} - \lambda\delta\bar{u}, \tag{7}$$

where $\delta\bar{u}$ and $\overline{v'q'}$ are to be interpreted as mass-weighted vertical averages over the height range 82–98 km.

Now, if we persist with the linear assumption and further assume that nonconservation of potential vorticity can be expressed by simple decay to a zonally uniform distribution at a rate λ_e, then it can be shown (e.g., Rhines, 1977) that

$$\overline{v'q'} = -\left(\frac{\partial}{\partial t} + 2\lambda_e\right)\left(\frac{1}{2}\overline{\eta^2}\right)\frac{\partial\bar{q}}{\partial y} \tag{8}$$

or, using (3) and assuming that $\partial\bar{q}/\partial y$ is dominated by the planetary contribution $\beta = 2\Omega$ cos (latitude)/a,

$$\overline{v'q'} = -\left(\frac{\partial}{\partial t} + 2\lambda_e\right)\left[A\exp\left(\frac{-t^2}{T^2}\right)\right] \tag{9}$$

where $A = \frac{1}{2}\beta|Y|^2$. The mass-weighted vertical average of A will be dominated by contributions from the lower levels, so, using $|Y| \approx 3000$ km, we find $A \approx 90$ m s^{-1}. (If, as noted above, $|Y|$ is actually larger at levels below 82 km, this may in fact be an underestimate of the total effective forcing.)

Using (9), it is straightforward to determine $\bar{u}(t)$ for a given λ. If the term $\partial(\overline{u'v'})/\partial y$ is negligible in (4) (it was argued previously that this term is small) then, with (7), we have

$$f\delta\bar{v} = -\overline{v'q'} \tag{10}$$

from which we may determine $\delta\bar{v}(t)$. Note that, for a given $\overline{v'q'}$, Eq. (10) implies that $\delta\bar{v}$ is independent of λ.

Thus far, we have not addressed the value of the mean and eddy decay rates λ and λ_e, except for the estimate $(0.5\text{ day})^{-1}$ based on the mean momentum balance. In general, λ_e will differ from λ, for at least three reasons:

(i) the effective frictional dissipation (of which the term $-\lambda\delta\bar{u}$ in (7) is but a crude parameterization) may act differently on the eddies than on the mean zonal flow;

(ii) the eddy motions will be influenced by thermal

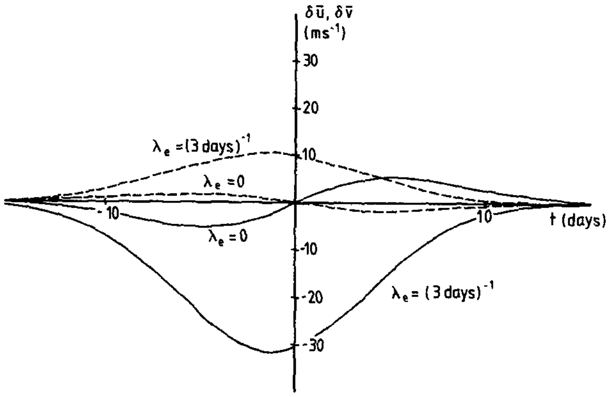

FIG. 6. Predicted response of mean flow (solid: $\delta\bar{u}$, dashed: $\delta\bar{v}$) to a wave pulse, calculated from (7), (9) and (10) with $\lambda_e = 0$ and $\lambda_e = (3\text{ days})^{-1}$.

as well as frictional dissipation which also may act differently on the eddies than on the mean, and

(iii) the formulation of simple decay of eddy potential vorticity may not be appropriate in any case.

Therefore, we regard λ_e as an unknown variable parameter. To reduce the number of variables, however, we fix the value of λ to be that estimated from angular momentum balance, viz, $(0.5\text{ day})^{-1}$. Given the simplicity of the present calculation, there seems little point in trying to refine this value further.

Results for a range of values of λ_e are shown in Fig. 6. When $\lambda_e = 0$ (conservative eddies), the response of the mean flow (in both u and v) is much weaker than observed and, contrary to the observed behavior, the response is antisymmetric about $t = 0$. The reasons for the antisymmetry are clear from (9)—$\overline{v'q'}$ is antisymmetric about $t = 0$ when $\lambda = 0$—and from (7) and (10), given that the time derivative in (7) is small (the timescale of mean flow dissipation, λ^{-1}, is small compared with the duration of the wave pulse). The introduction of eddy dissipation not only removes this asymmetry but also, from (9), intensifies the eddy forcing of the mean flow. Results for $\lambda_e = (3\text{ days})^{-1}$, also shown in Fig. 6, illustrate this and, in fact, these results agree reasonably well with the observed behavior of the "mean" circulation, with a maximum change of about 30 $m \cdot s^{-1}$ westward and about 10 $m\ s^{-1}$ northward, preceding the peak of the pulse by about 2 days (less than in the observations). It is, in fact, possible to improve agreement with observation further by finer tuning of the dissipation rates; given the crudity of the calculation, however, such an exercise would be meaningless. The object of this exercise is to demonstrate that, given plausible assumptions, it is reasonable to postulate that the observed changes in the low-passed wind field were a response of the mesospheric circulation to the two-day wave pulse.

4. Discussion

In most respects, the wave event described here was similar to previous manifestations of the quasi-two-day wave; in particular, the large amplitudes reached at the peak of the event are typical of earlier Adelaide observations. The pulselike nature of the event has also been noted in previous occurrences (Craig et al., 1983). The period of the wave was again significantly shorter than that (52 h) found in northern summer observations; however, while the period was found to be indistinguishable from 48 h in earlier years at Adelaide, it was close to 46 h in the January 1984 event.

It has been argued that the substantial changes occurring in the background flow in the mesosphere at the time of the wave pulse were, in fact, a direct consequence of transport by the wave. Actually, these changes were reasonably well reproduced by those predicted theoretically from the observed wave characteristics, provided it was assumed that

(i) the Adelaide observations are typical of all longitudes;

(ii) the factors (principally gravity wave transports) tending to restore the mean flow toward its climatological structure may be represented by Rayleigh friction at a rate of about $(0.5\text{ day})^{-1}$; and

(iii) the wave transport is nonconservative; this effect was represented by assuming simple decay of wave potential vorticity on a time scale of 3 days.

The first assumption, which was necessary to pursue such an analysis using data from a single location, appears to be consistent with what we know of the quasi-two-day wave. The second and third were suggested by the observations, in light of the analysis. Inclusion in the calculation of a restoration of the mean winds on a time scale of 0.5 days or so was dictated by consideration of the momentum budget; this is somewhat smaller than earlier theoretical predictions (e.g., Miyahara 1985). In response to a conservative, reversible wave pulse, the theory predicts a weak, induced flow which, in the decay stage of the pulse, is a reversal of that generated during the pulse amplification. The large qualitative and quantitative disagreement with the observed low-pass wind behavior suggests that the wave transport processes are nonconservative; agreement with observation was much improved by choosing an appropriate eddy dissipation rate.

Of course, the representation used here of wave decay processes is, at best, a crude parameterization of what may be a multitude of complex processes (e.g.,, radiative losses, gravity wave drag, turbulent mixing). In addition to these effects, there is also the possibility that the 2-day wave itself breaks in the upper mesosphere as planetary waves in the winter stratosphere appear to do (McIntyre and Palmer, 1983, 1984). If the estimates of parcel displacement amplitudes given here are actually realized, it would appear quite possible that the wave overturns in the horizontal plane, bringing about substantial north–south material transport. It is difficult to establish whether this occurs on the basis of the limited available information; one point of interest in Fig. 2, however, is the apparent change of character of the disturbance in the decay stage of the pulse (when the zonal wind component is much weaker than in the growth stage).

The mean wind changes are apparently limited both in magnitude and duration by influences restoring the climatological circulation. However, constituent distributions will also be influenced by these events. Indeed, given the large meridional displacements coupled with the possibility of wave breaking, the response of longer-lived constituents (i.e., with lifetimes in excess of 0.5 days) might be expected to be even more dramatic than that of the mean circulation. Transport by the quasi-two-day wave may be the most effective latitudinal transport process in the summer mesosphere.

Finally, it should be noted that the question of the origin of this phenomenon has not been addressed here. The foregoing analysis, as described, depends only on the observed wind structure of the wave and mean flow, and the wave period. There was, in fact, an implicit assumption made that the wave is not generated locally within the height range (about 80–100 km) of interest, but this seems reasonable, given the observed systematic vertical propagation in this region. Conversely, therefore, these results do not provide any clue as to the causality of the wave; certainly it is not possible, on the basis of the wave transport characteristics in the upper mesosphere, to distinguish a normal mode disturbance (Salby, 1981, 1984) from one generated by baroclinic instability of the lower mesospheric jet (Plumb, 1983; Pfister, 1985).

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