This suggests that the basic idea underlying the notion of a cutoff could in principle be more precisely defined in terms of IPV distributions. As far as we can see, such a definition would be entirely consistent with the traditional synoptic view of the essential phenomenon, taking other evidence into account such as wind and temperature fields (cf. Fig. 15) together with observational case studies of the time development of the cutting-off process. For instance section 10.1 of Palmén and Newton (1969, p. 274) describes the birth of a large cutoff cyclone from the cold "polar-source region", with which, at a certain stage of development, "it is still united by an 'umbilical cord' in the form of a shear line". From the information presented it appears that the stage of development referred to is fundamentally similar to that shown in our Fig. 5 for 23 September 1982, even though the orientation, geographical location, and other details are different.

Palmén and Newton describe the polar-source region as 'tropospheric' (*loc. cit.*, and top of p. 284). However, it has become increasingly clear, both from examples like that of Fig. 5 and from the theoretical principles reviewed in this paper, that this concept requires modification if one is interested in questions of dynamical cause and effect. For dynamical purposes an important part of the polar-source region is Kleinschmidt's lower-stratospheric reservoir of high-PV air. At least in cases like that of Fig. 5, the observed development appears to be largely controlled by long-range, quasi-isentropic advection of high-PV air from the lower-stratospheric reservoir. The word 'controlled' is used deliberately here, its use being justified by the invertibility principle. Whereas low temperature advection, for instance, may well appear important from a purely diagnostic point of view in, say, the middle troposphere ahead of the moving IPV anomaly, it can be argued that in terms of cause and effect its importance is actually secondary, in such cases, by comparison with that of IPV advection at higher altitudes. This is because much of the coldness of the free atmosphere beneath the IPV anomaly is attributable to the induced temperature field of the anomaly. As such, it *cannot be advected anywhere* unless

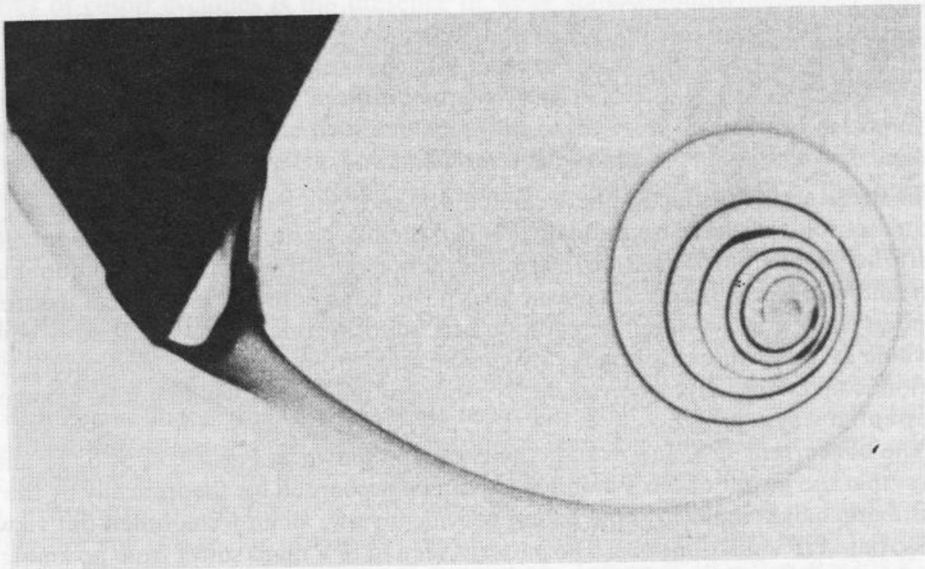

Figure 22. A standard fluid-dynamical experiment showing barotropic vortex rollup visualized by dye injection (Pullin and Perry 1980). A piston, not shown, drives water from left to right with almost constant speed normal to the axis of a wedge of 30° semi-vertex angle; the wedge acts as a source both of dye and of vorticity.

the anomaly is itself moving in the direction concerned. The point made in section 4 about the compensating effects of vertical motion is relevant here. Temperature advection near the ground is, of course, an entirely different matter, a fact which is related to the constraining effect of the earth's surface upon large-scale vertical motion and which has already been illustrated in several ways in section 5.

The phenomenon of cutting off exemplified in Figs. 5 and 11 appears to have counterparts in classical aerodynamics. Figure 22 shows what appears to be an aerodynamical (barotropic) counterpart to the 24 September panel in Fig. 5, as far as the cutoff cyclone and its presumed 'umbilical cord' are concerned. The figure, taken from a paper by Pullin and Perry (1980), represents a laboratory photograph using the dye method of flow visualization; the dye roughly marks high vorticity values (although the vorticity diffuses faster than the dye). The spatial resolution is, of course, far greater than that of Fig. 5, and the 'umbilical cord' shows up clearly. This type of flow is known to be accurately described by the barotropic vorticity equation, with a diffusive term included.

Features common to the aerodynamical and meteorological cases are the existence of a source of cyclonic (potential) vorticity fluid on the left, advection of cyclonic (potential) vorticity fluid from left to right, and a tendency for the furthest part of the (potential) vorticity distribution to *wind itself up* (a concept justifiable in terms of the concept of 'induced velocity field') into a compact, nearly axisymmetric vortex. Aerodynamicists use the term 'vortex rollup' to describe the phenomenon, and it has been extensively studied; see, e.g., p. 590 of the textbook by Batchelor (1967), and for more detail the review by Saffman and Baker (1979). The main difference between the two cases lies in the nature of the source region, which in the laboratory case is the boundary layer on a solid, wedge-shaped obstacle, seen at the left of the photograph, but in the meteorological case is Kleinschmidt's stratospheric reservoir of high-PV air. Also, in the atmospheric case there may well be much less spiral fine-structure in the IPV distribution than Fig. 22 might suggest, because of the different initial conditions. (And even if such structure were initially present—as was suggested in section 2(d) for another case—it would tend to be destroyed by small-scale quasi-barotropic shear instabilities.)

9. Concluding remarks

Perhaps the central point we have tried to bring out in this paper is the way in which the IPV concept succinctly encapsulates all the balanced dynamics usually described in terms of advection, divergence and vertical motion. IPV thinking gives direct insight, for example, into the circumstances in which the effects of advection and vertical motion tend to cancel each other; recall again the thought-experiment described in section 4. Especially for quasi-conservative processes involving rapid advection of upper air synoptic-scale features (sections 2(c), 6(e)), IPV thinking has considerable potential for furthering our understanding of the behaviour of real weather systems. Moreover, the invertibility principle suggests that the IPV concept should remain useful even in the presence of moist or dry diabatic heating or cooling (sections 6(e), 7), along with other non-conservative effects such as friction and gravity-wave drag. As was pointed out in section 4 the crucial advantage of IPV maps over, say, isobaric absolute vorticity maps, is the conceptual separation they offer between the effects of advection on the one hand, and the effects of vertical motion on the other.

The use of coarse-grain IPV maps together with surface θ maps should lead not only to a clearer recognition of partial analogies with extensively studied aerodynamical phenomena (section 8), but also to a greatly sharpened ability to relate observed phenomena to theoretical concepts generally. In order to make a meaningful comparison between dynamical theory, modelling and observation, it is often necessary to know something about IPV and surface θ gradients. For instance, such information may be necessary in order to tell whether or not a given instability mechanism could be operating, or could be about to operate, in a given synoptic situation (section 6).

The use of IPV maps could prove valuable as an aid to the quality control of numerical weather analysis and prediction, especially if appropriate image-processing and animated-graphics techniques were brought to bear. IPV maps might also be valuable in the development and assessment of numerical forecasting models, for example in problems like the embedding of fine-mesh regional models within larger-scale models. Fundamental to the success of such embedding is a sufficiently good representation, whether implicit or explicit, of the inflow of IPV features across the boundaries. This may be quite crucial, for instance, to predicting certain cases of explosive cyclogenesis (section 6(*e*)).

Similarly, it can be argued that the effectiveness of parametrization schemes for the physical processes governing diabatic and frictional changes in an atmospheric model (where 'frictional' refers to any sub-grid-scale momentum transfer process) should be judged, for dynamical purposes, largely by whether or not those schemes deliver the correct diabatic and frictional rates of change of IPV distributions when their output is substituted into the right-hand sides of Eqs. (70a) ff. To the extent that the invertibility principle holds, it tells us that the *only* dynamically relevant output of the 'physics package' used in the model is the net effect on the IPV distributions and on the low-level θ distribution, assuming of course that the package does not generate spurious gravity modes. The letter 'I' is important here, as elsewhere: a consideration of the accuracy of the rates of change of PV following an air parcel is not enough for this purpose, since IPV distributions can be greatly affected, also, by diabatic motion across isentropic surfaces when vertical PV gradients are strong, as has often been pointed out. This latter effect is allowed for by Eqs. (74a, b), whose right-hand sides remind us of the importance of having estimates of moist-convective and other contributions to diabatic heating with the correct vertical as well as horizontal distribution.

Another major area in which IPV maps could prove useful is that of research into tropical, extra-tropical interactions, in which a fundamental problem is to quantify the long-range interconnections between dynamical regimes having very different scaling characteristics. There are many aspects to be considered, of which the presence of subtropical upper air cutoff cyclones, such as that illustrated in Fig. 10, is merely one indication.

The 350 K IPV map shown in Fig. 2(a) and those for subsequent days (not shown) give numerous indications of low-PV air being injected into middle and high latitudes and high-PV air moving into subtropical and tropical latitudes. This is exactly what happens in the stratospheric 'breaking Rossby waves' referred to in section 6(*d*). As was also noted there, very similar phenomena occur in model simulations of nonlinear baroclinic-wave life cycles. The similarity of the trailing troughs or shear lines seen in Fig. 20(b) to that in the bottom left-hand quadrant of Fig. 2(a) is striking, as is the similarity of both patterns to those seen in the winter stratosphere; see also, e.g., Fig. 2 of Elliott (1956), Figs. 2, 5c of Holopainen and Rontu (1981), and Fig. 10.4a of Palmén and Newton (1969). One point of interest is that the characteristic spatio–temporal structure of the phenomenon, when viewed as a dynamical whole, makes it a likely candidate for helping to explain the interesting lagged correlation patterns between tropical cloudiness (cf. section 4) and mid-latitude 500 mb height field variability, recently discovered by Liebmann and Hartmann (1984).

The presence of features like D″ in Fig. 2(a) suggests that some of the high-PV air in the trailing trough may wind itself up into compact vortices in the manner suggested by the aerodynamical analogy discussed in section 8. It seems very likely that this is how subtropical cutoff cyclones of the kind illustrated in Fig. 10 are formed. The same thing appears to take place on a larger scale in the middle stratosphere (Clough *et al.* 1985; McIntyre and Palmer 1984), as well as in some of the other cases mentioned. The idea seems generally consistent with synoptic as well as aerodynamical experience and it would be of interest to repeat the numerical experiment of Fig. 20 at higher resolution, and to carry out other, related simulations, to see whether such behaviour is reproduced there as well.

Upper air trailing troughs and associated phenomena are unlikely to be the only significant features in tropical IPV distributions. For instance important IPV anomalies may be expected to be generated by large-scale diabatic heating. These will be subject to the integral constraint (70b).

The induced fields of any large-scale IPV anomaly, whatever its origin, may extend into, and thus affect, middle latitudes. As a result, the tropics may appear variously as a source, an absorber, or a reflector of mid-latitude planetary-scale disturbances. Thus, for example, the Rossby wave breaking phenomenon exemplified in Figs. 2(a) and 20(b) may represent a *two-way* interaction between middle latitudes and the tropics, not only injecting disturbances into the tropics, but also changing from day to day the extent to which the tropics appears to middle latitudes as an absorber or reflector of mid-latitude planetary-wave activity. The absorbing or reflecting characteristics of the tropics depend on the phase relations between mid-latitude and tropical IPV patterns in essentially the same manner as indicated by the discussion in section 6(*c*) and Fig. 19. This particular two-way interaction has been studied quantitatively, in idealized form, in the theory of 'nonlinear critical layers' referred to in section 6(*d*). Tests of the relevance of such theoretical ideas, and of many other ideas about the dynamics of tropical, extra-tropical interactions, will ultimately rest on knowing enough, implicitly or explicitly, about real IPV distributions.

Within the tropics one has the alternative possibility of using upper air vorticity diagnoses (e.g. Sardeshmukh and Hoskins 1985, and refs.), which will be locally equivalent to IPV diagnoses if the disturbances are sufficiently deep (in comparison with the appropriate Rossby height, Eq. (33b)), and if static stability values are sufficiently near constant. It is in describing dynamical interactions between regimes on either side of the subtropical jet, with its very steep static stability change, that IPV diagnostics seem likely to have a clear advantage.

The presence of negative PV regions in Fig. 2(a), if they are real, is suggestive of dynamically significant cross-equatorial advective processes. It should be noted in this connection that if an IPV anomaly in the form of an isolated vortex having a near-circular planform were to cross the equator into the opposite hemisphere, then the vortex would still be a stable entity except perhaps to a small extent on its periphery, depending on the ambient values of the potential vorticity P and hence of the product $f_{\mathrm{loc}}P$ in the inequality (30). It is not true, as is sometimes assumed, that the whole vortex would become inertially unstable since, as (21) and (30) remind us, it is the sign of f_{loc} and not that of f which is dynamically relevant in relation to the sign of the PV itself.

Many other applications of the IPV concept suggest themselves. Indeed the invertibility principle implies that there are potentially as many significant applications of IPV thinking as there are balanced dynamical processes of meteorological interest, whether linear or nonlinear, large or small scale. For instance the partial analogy with two-dimensional, barotropic aerodynamics immediately explains why we should expect to see Kármán vortex streets in the lee of mountainous islands, in the presence of stable stratification (e.g. Bugaev 1973; Thompson *et al.* 1977, and refs.), and also suggests ways of making the vortex-street concept quantitative, should this be desired, even though neither quasi-geostrophic theory nor any of its refinements, nor classical aerodynamics itself, is applicable. However, the most exciting prospects at present seem to lie in the study of synoptic and larger scales, where a significant amount of information is already available from operational analyses and forecasts, as has been illustrated here. In the present state of the art not all the features appearing in operationally-based IPV maps can be considered meaningful, of course, but it seems reasonable to hope for progressive improvement, especially if isentropic analysis methods are developed to their full potential. One incentive to progress is the fact that the quality of the best available operational IPV maps can be regarded as a sensitive measure of the quality of the analysis–forecast process itself.

In the past, practical application of the IPV concept to atmospheric data has always been rendered problematical by deficiencies in the data and by the volume of computation required. But now, in virtue of the quality already being attained in operational data analyses, and the availability of adequate computing power, coarse-grain IPV maps are beginning to look more and more like an extremely useful addition to the armoury of those interested in understanding and forecasting the behaviour of the atmosphere.

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APPENDIX

The computation of vertical motion

Various forms of the adiabatic, frictionless omega equation are available in the literature. We first derive a new form of the equation which is consistent with the IPV approach, beginning with the frictionless, adiabatic case.

As implied by (42), (43) and (44), the quasi-geostrophic potential vorticity q may be written as

$$q = f + \nabla_h^2 \psi' + f_0^2 \frac{\partial}{\partial p}(\mathcal{N}^{-2} \partial \psi'/\partial p). \tag{A.1}$$

Taking $f_0^2 \partial^2/\partial p \partial t$ of this equation and using the hydrostatic equation (41) and the definition of $\mathcal{N}^2$ (45) yields

$$f_0 \frac{\partial^2 q}{\partial p \partial t} = - R \nabla_h^2 \frac{\partial \theta'}{\partial t} - f_0^2 \frac{\partial^2}{\partial p^2}\left(\frac{\partial \theta'/\partial t}{-d\theta_{\text{ref}}/dp}\right), \tag{A.2}$$

where $R = R(p)$ is defined in (20). Following Andrews and McIntyre (1976) we define a 'residual' vertical velocity

$$\tilde{\omega} = \omega + (\mathbf{v}\cdot\nabla_p\theta')/(d\theta_{\mathrm{ref}}/dp) \simeq \omega + \nabla_p\cdot(\mathbf{v}\theta')/(d\theta_{\mathrm{ref}}/dp). \tag{A.3}$$

From the adiabatic thermodynamic equation, (A.3) may be rewritten

$$\tilde{\omega} = -(\partial\theta'/\partial t)/(d\theta_{\mathrm{ref}}/dp). \tag{A.4}$$

Using (A.4) to substitute for $\partial\theta'/\partial t$ in (A.2) gives

$$f_0\frac{\partial}{\partial p}\frac{\partial q}{\partial t} = -\mathcal{L}'(\tilde{\omega}), \tag{A.5}$$

where the linear operator $\mathcal{L}'$ is defined as

$$\mathcal{L}'(\omega) = \mathcal{N}^2\nabla_h^2\omega + f_0^2\partial^2\omega/\partial p^2. \tag{A.6}$$

The conservation of q moving with the geostrophic velocity enables the left-hand side of (A.5) to be written in terms of the advection of q and so the equation becomes

$$\mathcal{L}'(\tilde{\omega}) = f_0\,\partial(\mathbf{v}\cdot\nabla_p q)/\partial p, \tag{A.7}$$

which is of the form (49). After some manipulation it can be verified that substitution of (A.3) into (A.7) yields the usual omega equation.

(A.3) and (A.7) imply that the total vertical velocity may be written as the sum of three contributions:

$$\omega = \omega_{\mathrm{PVA}} + \omega_{\mathrm{BTA}} + \omega_{\mathrm{IU}} \tag{A.8}$$

where ω_{PVA} (potential vorticity advection) is the solution of

$$\mathcal{L}'(\omega) = f_0\partial(\mathbf{v}\cdot\nabla_p q)/\partial p \quad \text{with} \quad \omega = 0 \quad \text{on} \quad p = 0, p_0; \tag{A.9}$$

ω_{BTA} (boundary temperature advection) is the solution of

$$\mathcal{L}'(\omega) = 0 \quad \text{with} \quad \omega = (\mathbf{v}\cdot\nabla_p\theta')/(d\theta_{\mathrm{ref}}/dp) \quad \text{on} \quad p = 0, p_0; \tag{A.10}$$

ω_{IU} (isentropic upgliding) is the vertical velocity given by

$$\omega = -(\mathbf{v}\cdot\nabla_p\theta')/(d\theta_{\mathrm{ref}}/dp) \tag{A.11}$$

cf. (A.3). Here the sum of ω_{PVA} and ω_{BTA} is equal to $\tilde{\omega}$, satisfying (A.7) with the boundary condition implied by (A.3).

If the boundary temperature distribution is incorporated into the interior PV distribution according to (47), then ω_{BTA} is absorbed into ω_{PVA}. It should be noted that in real situations there can be cancellation between ω_{PVA}, ω_{BTA} and ω_{IU}, and that this cancellation is reference-frame dependent. Of course, in a frame moving with a steady system, the situation first considered in section 4, ω_{PVA} and ω_{BTA} are zero and ω_{IU} accounts for the entire vertical motion. The more general form derived here may be useful in suggesting qualitative corrections to simple isentropic relative flow analyses of vertical motion in systems which are changing with time.

For comparison we now note two other forms of the adiabatic, frictionless omega equation. In either case the boundary conditions $\omega = 0$ at $p = 0, p_0$ may be applied. The traditional form of the omega equation may be written

$$\mathcal{L}'(\omega) = f_0\frac{\partial}{\partial p}\{\mathbf{v}\cdot\nabla_p[f + \mathbf{k}\cdot(\nabla_p\times\mathbf{v})]\} + R\nabla^2(\mathbf{v}\cdot\nabla_p\theta'). \tag{A.12}$$

The two terms on the right-hand side are the vorticity advection and thermal advection terms, respectively. Following Hoskins *et al.* (1978) and Hoskins (1982), another form of the equation may be written

$$\mathscr{L}'(\omega) = 2\nabla_p \cdot \mathbf{Q} + \beta f_0\, \partial v/\partial p, \tag{A.13}$$

where the quasi-horizontal vector

$$\mathbf{Q} = -\,|R\nabla_p \theta'|\mathbf{k} \times \partial \mathbf{v}/\partial s \tag{A.14}$$

and s is a quasi-eastward horizontal coordinate in the direction of the θ' contour. It may be verified, again after some manipulation, that (A.12) and (A.13) are both mathematically equivalent to (A.8)–(A.11).

As discussed in Hoskins *et al.* (1978), the main reservation about the traditional vorticity-advection, thermal-advection form, (A.12), is the cancellation between the two forcing terms, the extent of which cancellation is frame dependent. The **Q** vector form (A.13), after some practice at determining the forcing using (A.14), allows some insight into the vertical velocity field given by quasi-geostrophic theory. The new form (A.8)–(A.11) is attractive in providing corrections to the simple isentropic upgliding associated with IPV and boundary θ advection relative to the system of interest and for linking the discussion with that of section 4. The sum (A.8) may on the other hand be too complicated for the simple diagnosis of other real situations.

In the presence of a frictional force-curl **K**, an Ekman boundary layer of height-scale δ, a diabatic potential temperature source $\dot{\theta}$, and topography of height h, ω has four additional contributions

$$\omega_{\mathrm{FRIC}}\ :\ \mathscr{L}'(\omega) = -f_0\partial(\mathbf{k}.\mathbf{K})/\partial p \text{ with } \omega = 0 \text{ on } p = p_0, 0; \tag{A.15}$$

$$\omega_{\mathrm{EKM}}\ :\ \mathscr{L}'(\omega) = 0 \text{ with } \omega = -\tfrac{1}{2}\rho g\delta\zeta \text{ on } p = p_0 \text{ and } \omega = 0 \text{ on } p = 0; \tag{A.16}$$

$$\omega_{\mathrm{DIAB}}\ :\ \mathscr{L}'(\omega) = -R\nabla_h^2\,\dot{\theta} \text{ with } \omega = 0 \text{ on } p = p_0, 0; \tag{A.17}$$

$$\omega_{\mathrm{TOPOG}}:\ \mathscr{L}'(\omega) = 0 \text{ with } \omega = -\rho_{\mathrm{ref}} g\mathbf{v}.\nabla_p h \text{ on } p = p_0 \text{ and } \omega = 0 \text{ on } p = 0. \tag{A.18}$$

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