A BALANCED FLOW PERSPECTIVE ON THE EFFECT OF DEEP CONVECTION

G.J. Shutts and M.E.B. Gray
U.K. Meteorological Office
Bracknell, England

Summary: Deep, penetrative convection can be viewed as a process that redistributes mass in a stratified fluid and, in the presence of background rotation, leads to the creation of potential vorticity anomalies and accompanying balanced flows. The kinetic energy released in convective updraughts may eventually be trapped in these mesoscale balanced motions and then undergo an upscale energy cascade. In the case of mesoscale convective systems, the direct impact of vertical mass transfer on synoptic flow evolution may be considerable. Here we examine the efficiency with which convective available potential energy (CAPE) is converted to mesoscale, balanced flow energy and speculate on the need to represent these processes in forecast models.

1. INTRODUCTION

The substance of this paper is concerned with a rather idealized view of deep convection which focusses on the relationship between the storm and its stratified environment. The details of storm updraught/downdraught organization and cloud microphysical processes are regarded as important only in as far as they affect the redistribution of air parcels to their final balanced positions. This has the virtue of concentrating on those issues relevant to the way a forecast model would respond to convective mass transfer – if it were represented as a singular rather statistical process.

We address some fundamental issues concerning the fate of kinetic energy released in convective updraughts. Most convection parametrization schemes implicitly assume that this energy is dissipated by turbulence within a model grid-column. Those schemes that handle vertical convective momentum transport usually assume that this process is downgradient and that – in the mean – kinetic energy is extracted from the vertical wind shear of the environment — again implying that the energy is dissipated by turbulence. On the other hand, horizontal variations of parametrized convective warming in a numerical weather prediction model (NWP) imply the generation of available potential energy and therefore kinetic energy through geostrophic adjustment.

Knowing that mesoscale convective systems can create quasi-balanced vortices whose energy may cascade upscale in the sense described by two-dimensional turbulence theory, it seems worthwhile investigating the efficiency of this energy trapping mechanism. This goal is made difficult by the diversity of forms that convection takes in atmospheric flows. It is expected that the efficiency of balanced flow production in an ensemble of cumulus clouds is much less than that for a mesoscale convective system. The efficiency may depend on the Coriolis parameter so that tropical convective clusters behave differently from mid-latitude storms. There is also the possibilty that slantwise convection in mid-latitude weather systems contributes to the production balanced flow energy. Although diagnoses of numerical model case studies of cyclogenesis often show the existence of slantwise convective available potential energy, it is not clear whether this represents a real source of unbalanced flow energy, or is a product of using a two-dimensional theory to diagnose three-dimensional flow energetics.

Satellite imagery, and in particular animation loops, provide some inspiration as to the variety of form and effects of deep convection. It also reminds us of how inappropriate the implementation of grid-column-based convective parametrization can be. Cold airstream convection may evolve from being a shallow boundary layer phenomenon to a clustering, on the scale of ten of kilometres, of deep convective cloud within the circulation of mature cold-core depressions. In summer over land, elliptical mesoscale cloud anvils can grow within hours to cover areas hundreds of kilometres across suggesting the release of much conditional instability. Other severe storms form at oceanic cold fronts and suggest a close cooperation between synoptic forcing and the sustained release of CAPE in boundary layer air. The imagery encourages us to believe that there are some forms of convection for which the upward transfer of mass is fairly undiluted by mixing.

Green et al (1966) hypothesized that moist, boundary layer air ascending as a quasi-steady 'conveyor belt' in a baroclinic wave system, satisfies a generalized form of the Bernoulli equation governing the speed of the outflow beneath the tropopause. Case study analysis of air trajectories originating in the Trade Wind boundary layer which ascended to the upper troposphere gave credence to this hypothesis. In effect they were proposing that the polar-front jet speed is determined by the amount of CAPE seen by sub-tropical air parcels with respect to some slantwise trajectory : no energy loss by turbulent dissipation or gravity wave radiation was assumed. This rather remarkable picture, based on energetics alone, can only be half the story of course because vorticity constraints are completely ignored.

A similar class of steady convection models developed by Moncrieff and Green (1972) also used a Bernoulli theorem to determine convective storm structure and propagation speed. In this type of model, the convection is not distinct from a rotating, stratified environment and the entire atmosphere is effectively overturned. Here again, no energy is considered to be dissipated, radiated as internal gravity waves, or used to generate balanced rotational flows. These types of model have been used to infer the transfer properties of mid-latitude and tropical convective storms with a view to formulating parametrization schemes for heat, momentum and moisture (Moncrieff, 1992). An obstacle to the realization of this goal is the requirement that a parametrization should represent the effect of an ensemble of convective clouds embedded in an environmental flow.

In this paper we will concentrate on those convective systems that are not usefully described as members of a cloud ensemble (e.g. mesoscale convective systems) yet make an important contribution to the sub-synoptic scale potential vorticity field. We shall consider some simple mass transfer models which highlight the production of balanced energy from kinetic energy released by CAPE.

2. IDEALIZED MODELS

The release of convective instability in deep penetrative convection is rarely symmetrical in the vertical and one usually expects to see a net upward flux of mass (i.e. the updraught mass transfer exceeds the downdraught transfer). In the simplest view of penetrative convection, the downdraught may be ignored and we may focus on the effect on the environment of transferring mass from the lower to upper troposphere in a *before*-and-*after* sense. A particularly appealing picture of this adjustment process was provided by the axisymmetric element model of Shutts et al (1988) which sought the end-state following the convection of a cylindrical region of air. Fig. 1 shows the initial state in this element model where each element represents the cross-section of a torus of fluid characterized by uniform potential temperature (θ) and angular momentum (M). Initially, elements in rows have the same θ and elements in columns have the same M, and this discretized state represents a rotating atmosphere at rest. The shaded elements are assumed to be saturated, unstable to vertical displacements, and have a neutral buoyancy level located at a height of 7 km. Mixing with the environment and precipitation are ignored in the adjustment. To find the endstate after these elements have convected, their potential temperature is set to the value at a height of 7 km and a new equilibrium state computed. In this fluid rearrangement, the angular momentum and mass of each torus are conserved exactly and the final state is one of gradient wind and hydrostatic balance.

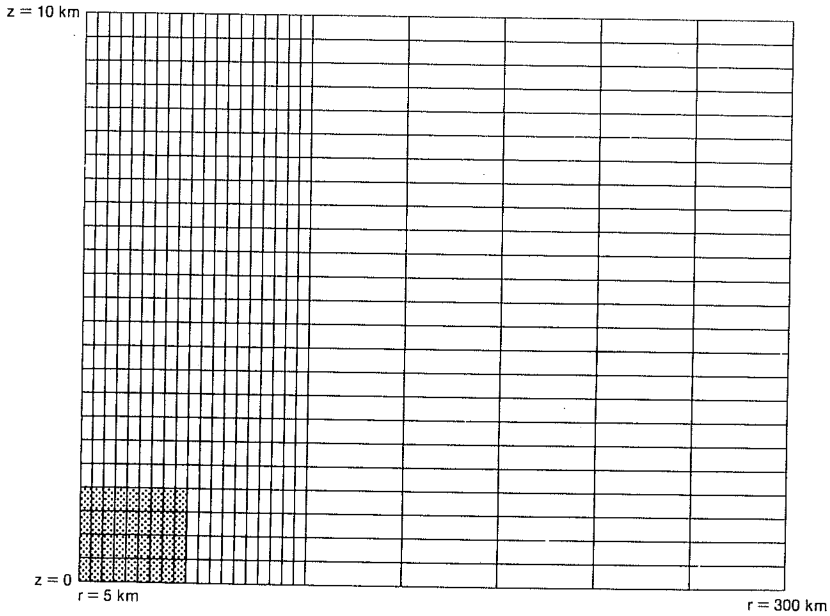


Figure 1: Initial state of Geometric model representing a regular piecewise constant discretization of the potential temperature and angular momentum fields. The shaded elements are those that are able to convect.

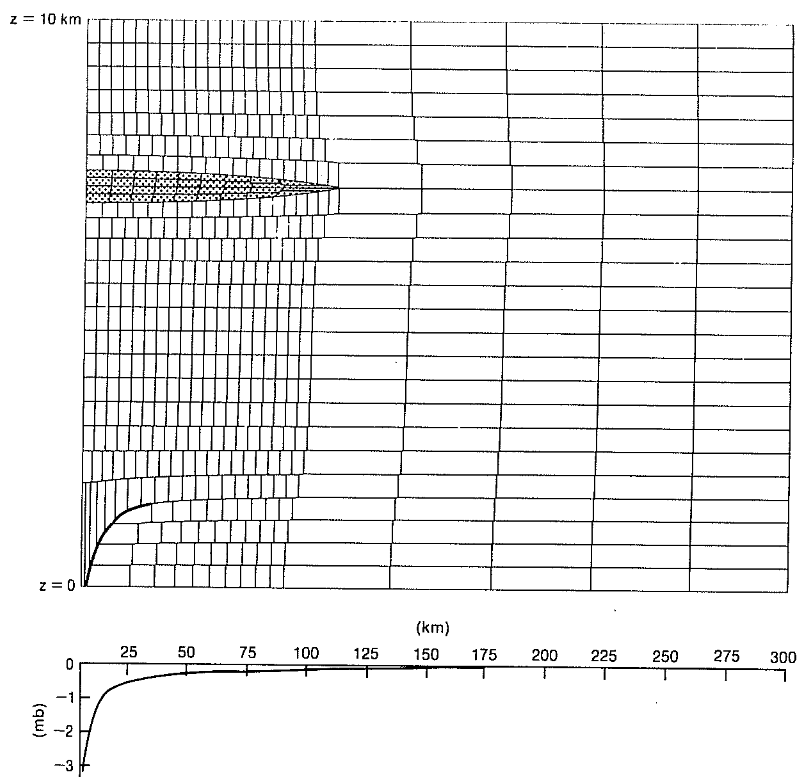


Figure 2: The element configuration after the shaded elements have convected. The thick line bounding elements at low-levels near the vortex core denotes a frontal boundary between air that has subsided in the core of the vortex and air from larger radius that has moved in to replace the shaded elements. The surface pressure is plotted beneath the element diagram.

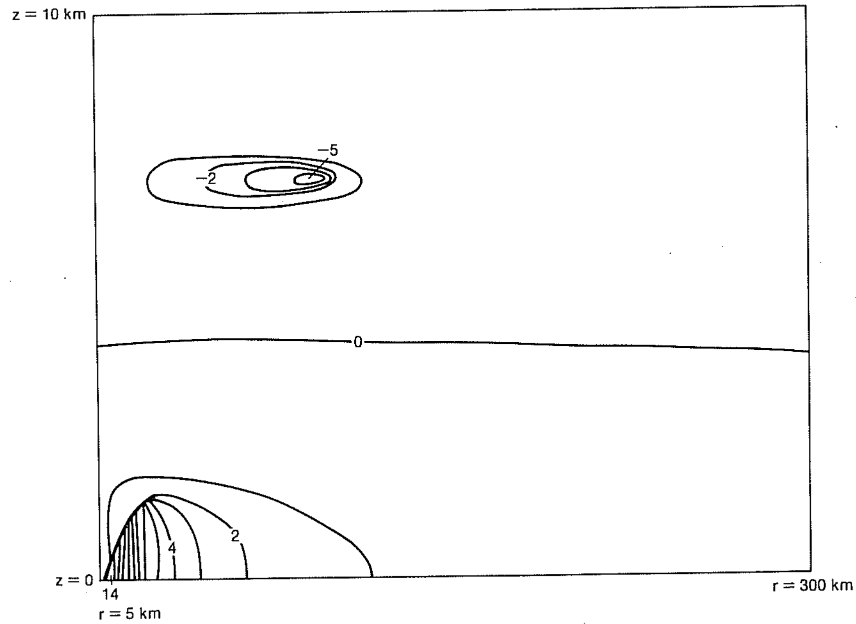


Figure 3: The azimuthal wind component (in m/s) inferred from the distribution of angular momentum in the final, balanced state.

Fig. 2 shows this new balanced state comprising an upper level lens of shaded fluid parcels and a low-level warm-cored vortex bounded by a frontal surface which slopes outward with height. The corresponding tangential wind field (Fig. 3) shows that the outer edge of the lens is accompanied by a shallow anticyclonic jet and at much smaller radius near the surface lies an intense cyclonic vortex. Beneath the lens and within the vortex, adiabatic warming is caused by the downward displacement of elements. Note however that forced ascent in the upper-half of the lens and above causes *cooling.*

This pair of balanced flow features are consistent with the creation of potential vorticity (PV) anomalies : a zero PV lens and frontal region of very high PV. It was conjectured by Shutts et al (1988) that in the limit of infinite element resolution, the front would represent a true discontinuity at which the PV was infinite. The process of PV forcing here can be related to the dilution and concentration of PV substance through diabatic mass transfer as proposed by Haynes and McIntyre (1987, 1990). Alternatively, one can suppose that scale of the convective updraught is too fine to resolve and the associated mass transfer leads to an apparent mass source and sink. Indeed, if Ertel's theorem is rederived with a hypothetical mass source term M_f in the continuity equation it becomes:

$$\frac{Dq}{Dt} = -qM_f \tag{1}$$

implying that mass sources lower the PV (defined here as q) and mass sinks increase it. Note that mass forcing cannot force negative q if it is everywhere positive initially.

This is consistent with the view that infinite dilution of PV substance gives zero q.

The production of balanced flow energy is caused by creation of well-separated and strong PV anomalies. Analytic models resembling this picture of convectively-forced lens and front have been obtained using the semi-geostrophic equations (Shutts, 1987; 1991; 1995). It can be shown that if $\hat{r}$ is the horizontal scale of a zero PV lens generated by deep convection and $\hat{E}$ represents the total balanced energy in the adjusted state then:

$$\hat{E} = \hat{r}^4(a + b\log h + O[h^{-1}]) \tag{2}$$

in two dimensions and

$$\hat{E} = c\hat{r}^5 \tag{3}$$

in three dimensions where a, b and c are constants, h is the vertical mass source-sink separation distance and the hat denotes a non-dimensional variable. If H is an arbitrary height scale, then NH and NH/f are used to non-dimensionalize velocity and horizontal length respectively. Since the mass convected M_c is proportional to r^2 in 2D and r^3 in 3D, $E \propto M_c^2$ in 2D and $E \propto M_c^{5/3}$ in 3D.

In the two-dimensional line source case, the energy associated with each opposite-signed PV feature is unbounded and only becomes finite when mass source and sinks of equal strength are considered together. Even then the total energy increases logarithmically as the source and sink PV features are pulled apart. However, a localized three-dimensional PV anomaly has bounded energy dependent on the 5/3 power of mass intruded or extruded to or from a rotating, stratified flow.

Now an upper limit to the amount of kinetic energy that may be released from the rate of working of the buoyancy force is $CAPE \times M_c$. Since this increases linearly with M_c but the work required to create the balanced flow depends on a higher power of M_c, there will be an upper bound on the amount of mass that may be transferred in a localized convective storm. In the 3D case, the dimensional balanced energy E can be shown to be:

$$E = c\frac{f^3 r^5}{N} \tag{4}$$

and the work done by buoyancy forces to convect the equivalent amount of mass scales as $CAPE \times 4\pi r^3/3(f/N)$.

In the (unattainable) limiting case where all of the energy released in convection is used to generate balanced flow, the above expressions may be equated to give the maximum radius of a convectively-generated lens r_* where

$$r_* = \beta \frac{\sqrt{(CAPE)}}{f} \tag{5}$$

and

$$\beta = \left(\frac{4\pi}{3c}\right)^{1/2}. \tag{6}$$

In the analytic model of Shutts (1991), β was found to be of order unity implying a maximum possible mesoscale cloud anvil of about 400 km for an environment with CAPE equal to 1600 Jkg^{-1} and $f = 10^{-4}$. Such expressions are unlikely to have any practical value since many important cloud processes have been ignored e.g. compensating downdraught mass fluxes, dissipation and the effects of precipitation. They do, however, provide a useful reference point in this balanced view of penetrative convection.

3. NUMERICAL SIMULATION OF THE ADJUSTMENT

In order to gauge whether these assumptions concerning the generation of balanced flow energy by convection hold, a series of numerical modelling experiments have been carried out by Gray (1996). The model is based on the anelastic, quasi-Boussinesq set described in Shutts and Gray (1994) and includes the water phase through liquid water temperature and total water mixing ratio variables. The thermodynamic equation is based on the conservation of moist static energy and, in principle, is somewhat less accurate than a scheme based on moist entropy conservation. The model domain is cyclic in the horizontal and uses a stretchable grid in all three spatial coordinates. This is employed to good effect in the horizontal so that extra resolution can be achieved in the vicinity of the deep convection.

In the 2D simulations to be described , 800 points are used in the horizontal with a spatial separation varying from 200 m in the middle to 8 km at the domain boundaries — 1000 km from the centre. The domain extends up to a height of 18 km of which the top 8 km is used as a damping layer for internal gravity waves. 180 levels are used in the vertical with a spacing of 77 m below 10 km smoothly varying to 160 m in the damping zone. The initial state potential temperature increases exponentially with height so that the buoyancy frequency N is given by $N^2 = 1.4 \times 10^{-4}\ s^{-1}$, the surface potential temperature is 291 K and the Coriolis parameter $f = 10^{-4}\ s^{-1}$.

In this environment, saturated air in the lowest few kilometres is convectively-unstable and on the basis of parcel theory would ascend to a height of about 6 km. Control is exerted over the volume of mass that convects by considering runs with differing horizontal extents of moist air. Both 2D and 3D simulations are initialized with a warm, cosine-squared, moist bubble of width $X_\theta = 2$ km, height H_θ=1.25 km , and perturbation amplitude of 0.5 K at the centre of the domain.The bubble is surrounded by a region of 99% relative humidity of diameter X_Q up to a height of H_Q. Outside of a central region of diameter X_B and height $H_Q = 5$ km, the relative humidity is set to a background value of 60% (Fig. 4). By varying the parameters X_Q and X_B the amount of mass available for convection can be controlled. X_θ was held fixed at 2 km.

Fig. 5 shows the component of wind normal to the vertical plane (i.e. parallel to the axis of convection) in a sequence of runs with increasing X_Q and X_B, after 50 Coriolis time periods (500,000 s). The flow is in geostrophic balance and comprises a vertical shearline front up to the base of a lens region where the absolute vorticity is close to zero. As expected the intensity of this flow is determined by the amount of mass that has convected.

Fig. 6 shows a plot of the total kinetic energy released during each simulation plotted against the total mass that was deemed to have convected. Apart from a single rogue point for the highest mass transfer, these points lie comfortably on a straight line demonstrating the existence of an effective CAPE for these experiments. Note however that the line does not pass through the origin.

Fig. 7 shows the corresponding balanced energy for each of these runs plotted against M_c^2. The good straight line fit of the points agrees with the expected power law between E and M_c for 2D convection given that h, the depth of the convection, is essentially constant in eq.(2).

A similar set of 3D runs were carried out with a horizontal grid comprising 80 points in both x and y (with gridlength varying between 200 m and 9 km) and 150 levels in the vertical, with spacing equal to 90 m below a height of 10 km. In these runs it was necessary to decrease the Rossby radius of deformation by a factor of ten to accommodate the balanced response within the smaller domain : this was achieved by using a Coriolis parameter $f = 10^{-3}\ s^{-1}$. It was also found necessary to initialize the model with several warm 'bubbles' rather than a single one as in the 2D runs to elicit a coherent MCS response. A full account of these experiments is given in Gray (1996). Fig. 8, showing the balanced flow energy after 50,000 s of integration plotted against $M_c^{5/3}$, appears to support the 5/3 power law inferred from the mass forcing model.

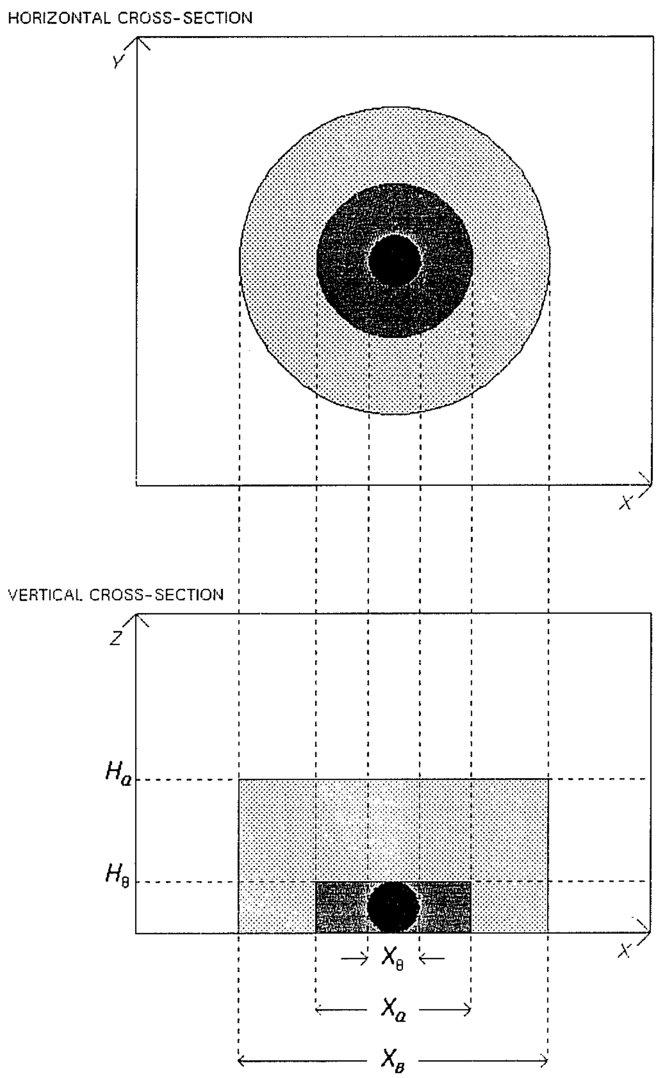


Figure 4: Diagram showing the layout of the initial θ and relative humidity fields for the plume experiments. The black shading identifies the warm bubble; the dark grey shading marks regions of 99% relative humidity, and the light grey shading marks the region where the relative humdity falls linearly to the background value of 60%.

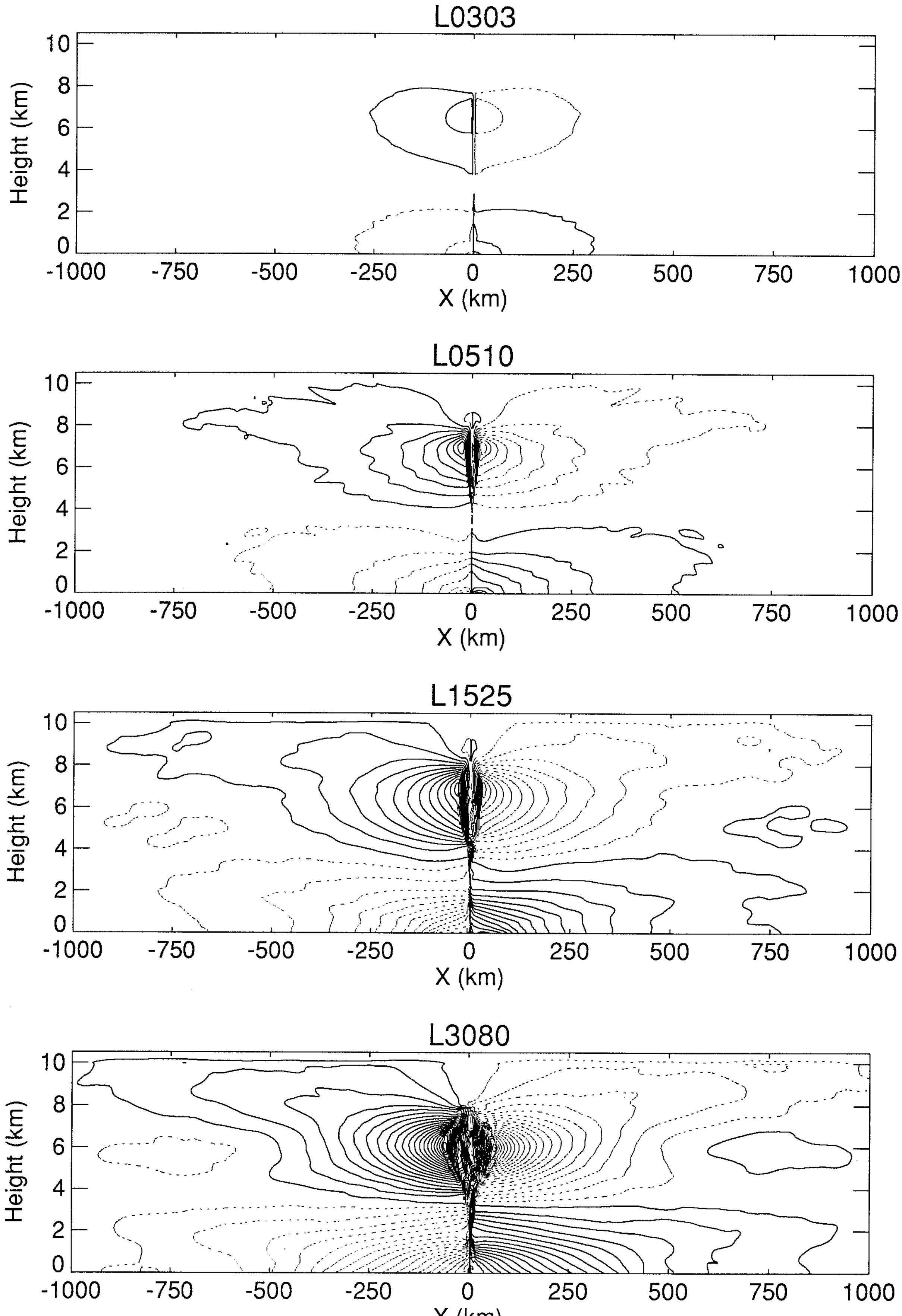


Figure 5: Wind component into the picture for increasing amounts of mass convected. Solid contours are into the picture and the interval is 0.25 ms^{-1}.

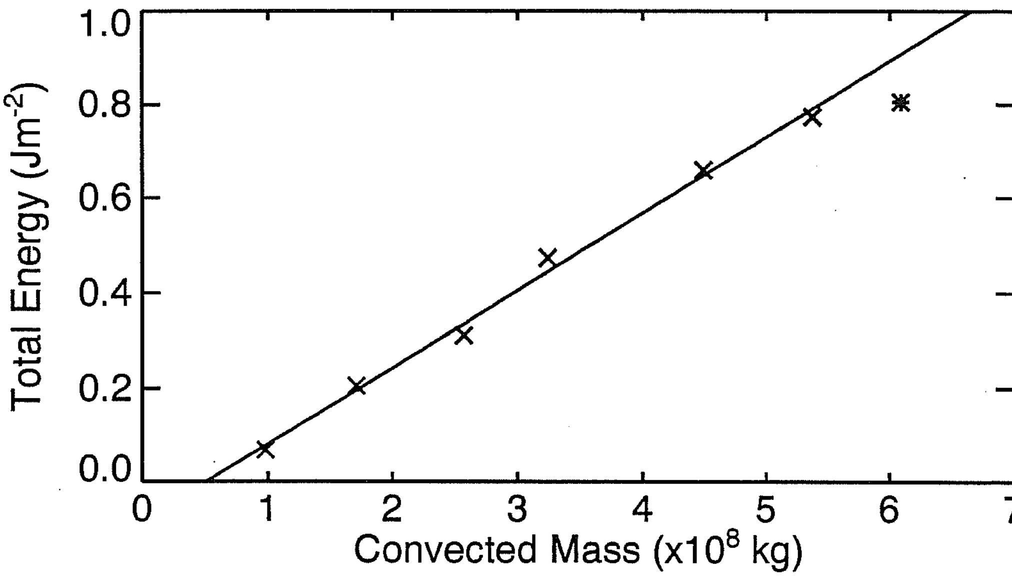


Figure 6: Total kinetic energy released by buoyancy forces for runs where differing amounts of mass convected.

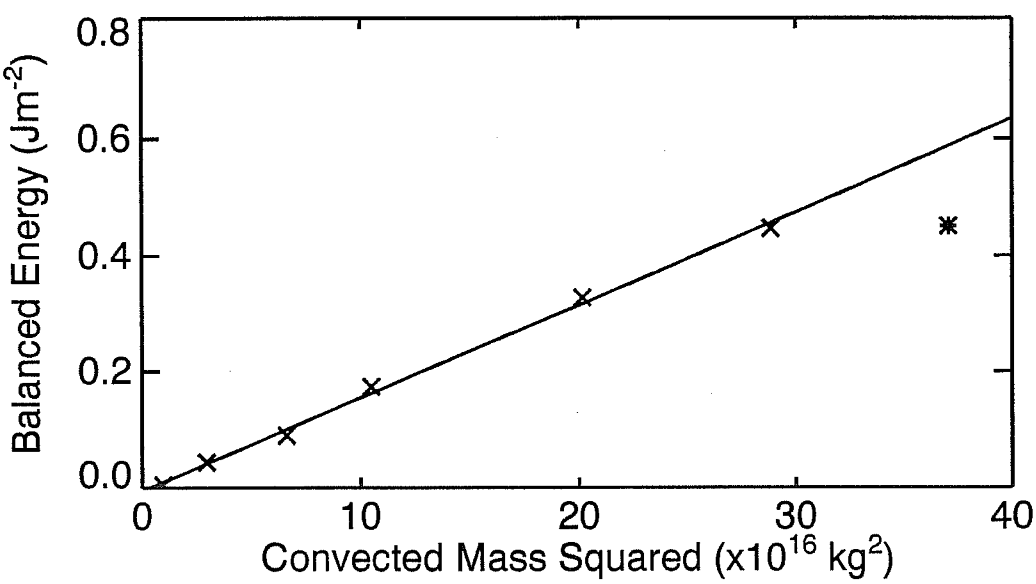
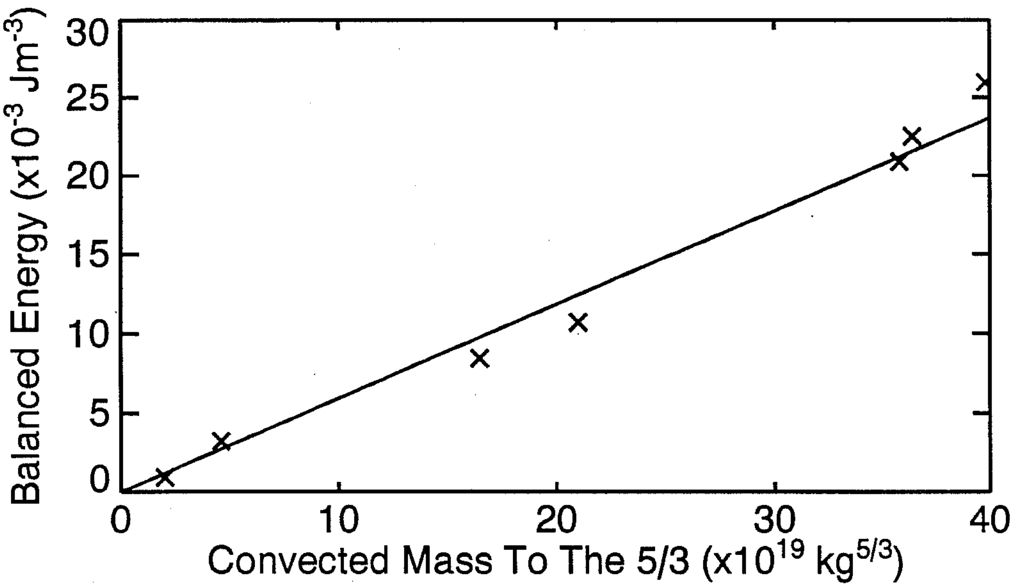


Figure 7: Residual balanced energy corresponding to the same 2D runs as in Fig. 6



In the 2D run giving the largest M_c value, about 57% of the energy released by convection was captured in balanced flow. This compares with a corresponding value of 38% in the 3D run where dissipation was much higher. Implicit numerical diffusion associated with the advection scheme is the main source of energy dissipation in the model with only a small contribution arising from the sub-grid turbulence model. For the 2D run about 1/6 of the kinetic energy produced is dissipated compared to almost half for the 3D run with the greatest mass flux. Gravity wave radiation accounts for the remaining part of the energy budget which for the 2D case is about 57 % compared to only 9% in the 3D case. What is striking about these figures is the high proportion of the energy released that is retained as balanced flow when a large region of moist boundary layer air is available to convect.

At this point it should be noted that the initial anvil cloud flow forced in the 3D runs is inertially-unstable and undergoes an axisymmetrization process. High potential vorticity generated by the effective mass sink in the convective updraught extends into the anvil lens region and is 'shielded' by a negative PV halo. The consequent decrease in angular momentum from the cyclonic core to the anticyclonic outer flow results in a non-symmetric inertial instability that re-establishes the monotonic increase of angular momentum with radius along isentropic surfaces.

Having established that the power law expression :

$$E = aM_c^n \tag{7}$$

(with $n = 2$ in 2D and $n = 5/3$ in 3D) holds reasonably well within the parameter space of the numerical experiments, we may now show that the efficiency of balanced flow energy generation is greater for a single MCS-type cloud system than for an ensemble of deep convective clouds carrying the same upward mass flux. If each cloud in an ensemble of N clouds convects mass $\hat{M}_c$, then assuming the adjustment process within each is independent, the total balanced energy created will be $Na\hat{M}_c$. However, if all of this total mass flux ($N\hat{M}_c$) occurred in a single MCS system then the balanced energy would be $a(N\hat{M}_c)^n$ and their ratio would be $1/N^{n-1}$. Therefore, if the exponent $n > 1$, the single convective system will be more efficient at generating balanced flow energy.

4. UPSCALE ENERGY CASCADE AND LILLY'S HYPOTHESIS

Many case studies of MCS events have demonstrated that anticyclonic flow forced in the upper troposphere has a direct impact on the synoptic-scale flow e.g. the formation of jet-streaks (Maddox, 1983; Cotton et al, 1989). This is consistent with the findings of the last section that about one half of the energy released in MCS-like storms is captured in balanced flow. However, for other forms of convection organization (e.g. deep convective cloud clusters in cold, cyclonic airstreams over warm seas) the efficiency will be much lower and the physical mechanism responsible for the generation of PV anomalies may be different. Even so, the larger areal extent of such convection may compensate for the poor efficiency of balanced flow production.

Lilly (1983) proposed that convective storm anvils might behave like collapsing turbulent wakes in stratified flows and that the residual vorticity normal to isentropic surfaces (i.e. associated with PV anomalies) would partake in an inverse energy cascade to synoptic scales. The hypothesis was inspired by aircraft observations of the horizontal energy spectrum (e.g. Gage and Nastrom, 1986) which showed a $-5/3$ power law in horizontal wavenumber extending from wavelengths of $\sim$ 1 to 600 km; and by Gage's proposal that the mesoscale energy spectrum is forced by upscale energy cascade from small-scale turbulence (Gage, 1979). The $-5/3$ power law was taken as possible evidence for an inertial-subrange of two-dimensional turbulence which he showed would be valid for small Froude number and large Rossby number.

The source of vorticity envisaged in this process is different from that discussed so far in this paper since Lilly did not consider background rotation (i.e. $f = 0$). In real situations, vertical vorticity may be derived from ambient vorticity or through baroclinic production combined with tilting into the vertical. As a first step in quantifying the mechanism proposed by Lilly, a series of numerical modelling experiments have been carried out by Vallis et al (1996). Using the same model as described in Section 2, they forced a ensemble of convective clouds by imposing constant cooling over a domain $300 \times 300 \times 12$ km given a fixed 'sea' temperature and appropriate surface heat and moisture transfer laws. A uniform background geostrophic wind was imposed through a fixed horizontal basic state pressure gradient.

The model integration was continued for 3.5 days and the evolution of the horizontal energy spectrum at the average cloud top height was observed.

Fig. 9 shows the horizontal energy spectrum split into rotational, divergent, vertical and APE components at a time 175000 s into the integration. The rotational energy spectrum is here associated with the balanced part of the flow (i.e. that part of the flow which will persist after the convection ceases). Energy released in convection has extended upscale in a manner reminiscent of the $k^{-5/3}$ inertial-subrange of 2D turbulence theory (k is the wavenumber magnitude) though this also holds for the divergent part of the flow. Internal gravity waves are trapped in this model domain and there is no damping layer and so a build-up of 'sloshing-modes' occurs. When the convection is finally suppressed by switching-off the cooling and surface energy fluxes, the divergent energy spectrum collapses leaving the rotational spectrum dominant.

Similar runs with (i) no Coriolis force (except on the horizontally-averaged flow) and (ii) fixed horizontally-averaged wind speed (effectively removing boundary layer wind shear) *and* no Coriolis parameter, were used as sensitivity experiments to examine the removal of ambient vorticity. Fig. 10 shows the rotational energy spectrum at time 170000 s in the control run together with these two sensitivity runs. It is apparent that whilst the absence of background rotation diminishes the energy in large scales, the biggest effect comes from removing the mean horizontal boundary layer vorticity. This should be contrasted with the squall-line simulations of Davis and Weisman (1994) in which baroclinically-generated vorticity was the ultimate source for the 'bookend' vortices at the extremities of the simulated squall line.

In another run, Vallis et al (1996) suppressed the effects of precipitation by removing all condensed water the instant it forms. This substantially increased the upscale energy cascade by eliminating the production of convective downdraughts – thereby increasing the net upward mass flux. In fact, it was found that 'mini-hurricane' vortices with diameters

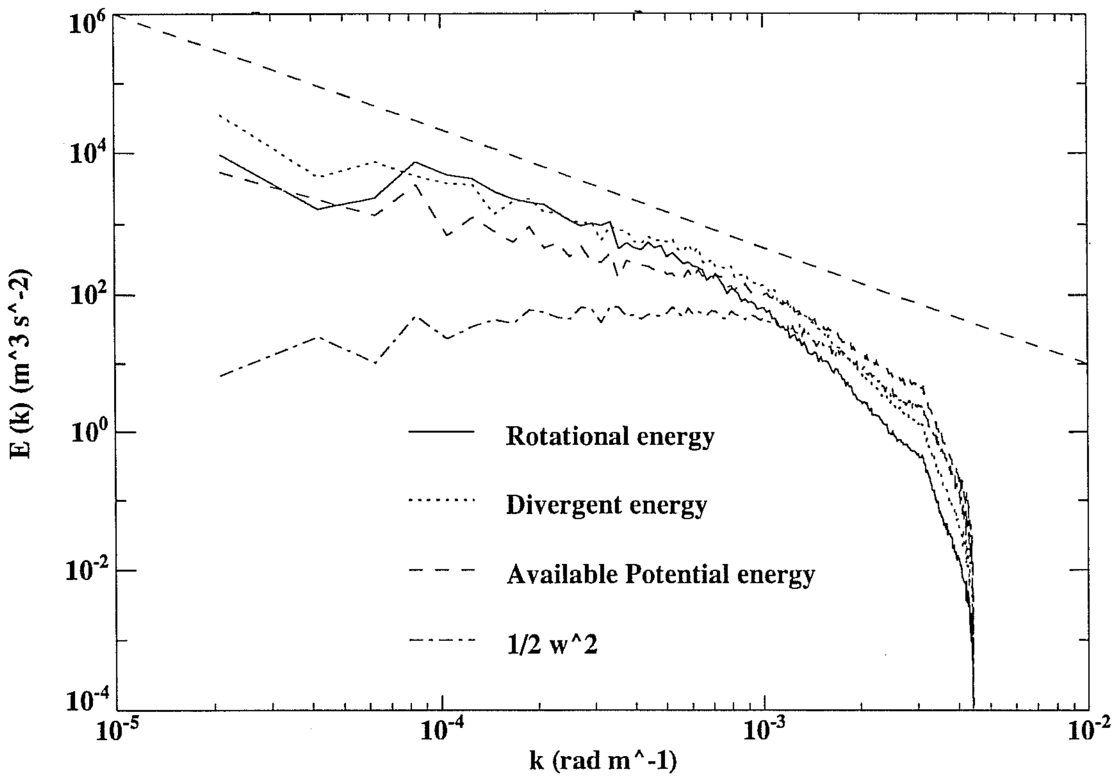


Figure 9: Components of the energy spectrum at time 170000 s.

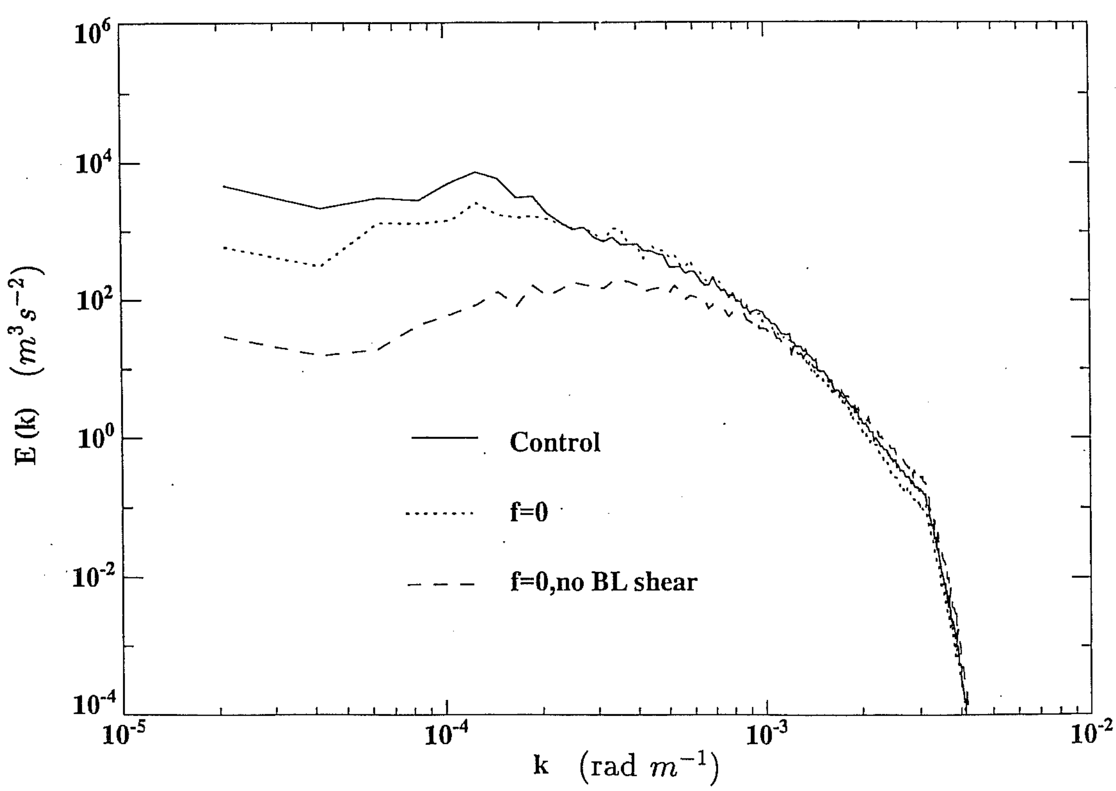


Figure 10: Comparsion of rotational energy spectra at time 170000 s for runs as described in text

of $\sim$ 10 km, azimuthal wind speeds of 22 ms^{-1} and warm-cores extending throughout most of the depth of the domain, were formed. This process is, in itself, quite interesting since hurricane modellers normally have to initialize their numerical simulations with a pre-existing vortex. It may be that in real situations where the atmosphere is nearly saturated through deep layers, downdraughts are not readily formed and the 'no precipitation' described here run is a crude model of this situation.

Vallis et al (1996) concluded that the upscale energy cascade was weak in their simulations and that this was related to the small mass flux occuring in these short-lived model clouds. Their simulated updraughts were essentially vertical and so no sustained mass transfer was possible in the presence of precipitation. The efficiency of the upscale energy cascade in regions of convective activity therefore remains an open question and is likely to depend on the large-scale flow environment (e.g. low-level moisture convergence) and cloud organization and microphysics.

5. DISCUSSION

This paper has raised some issues concerning the adjustment of a rotating, stratified fluid to convective mass transfer. From this before-and-after perspective, convection acts as a distributed mass source or sink of environmental air causing the concentration and dilution of potential vorticity substance. In the simplest picture, a non-entraining updraught drawing air from the boundary layer provides a mass source at its neutral buoyancy level and a mass sink in the boundary layer itself. These in turn force a zero PV lens and surface warm-cored cyclonic vortex. If evaporational cooling due to precipitation (and melting of ice) generates a convective downdraught from middle-levels to the boundary layer, the net mass forcing in the boundary layer may change to a mass source and a mass sink would appear at middle levels. In this case the surface warm-core cyclone would be replaced by a cold anticyclonic dome and a cyclonic vortex would appear at middle-levels (Raymond and Jiang, 1990). Mixing processes also contribute to the effective mass forcing by entrainment or detrainment. For instance, environment air may be entrained into the convective updraught and carried upward into the cloud anvil : this ingestion of mass at the updraught forces positive PV locally.

The idea of introducing mass sources or sinks as fictitious terms in the continuity equation might at first sight seem unwise. It is perhaps the only equation in NWP models which conventionally does not contain 'physics parametrization'. Certainly, notions of the molecular diffusion of density are excluded from the continuity equation by defining the concept of fluid velocity appropriately. On the other hand, precipitation represents a real sink of mass in saturated air parcels and in other planetary atmospheres the loss of mass due to gaseous phase changes could be important on dynamical time-scales . In laboratory experiments using rotating, stratified fluids, mass withdrawal/intrusion is a commonly-used technique for generating balanced vortices (e.g. Hedstrom and Armi, 1988). Given the direct relationship between mass forcing and PV forcing it may be a useful technique for modifying the potential vorticity field in NWP models. The practicality of this approach has been clearly demonstrated by Gray (1996) in a Large Eddy Simulation model context.

Convection parametrization schemes based on a mass flux model would appear to provide the basis for explicit mass forcing as an alternative to 'apparent' subsidence warming and drying of conventional schemes. In a mass forcing scheme, air parcels in an NWP model would undergo forced descent rather than receive diabatic heating in-situ. Above the height where most of the mass flux detrains, an explicit mass flux scheme would directly force adiabatic cooling whereas a conventional parametrization scheme would only obtain the cooling in the adjustment that follows the parametrized convective warming. Similarly, the lateral spreading of air at cloud top level would be directly forced with an explicit mass forcing parametrization. These apparent advantages in a mass forcing approach may, however, be lost for an ensemble of convective clouds – i.e. the conditions under which convective parametrization is justified.

For many years, the previous version of the UK Meteorological Office operational mesoscale model used a 'parametrization' of individual thunderstorms based on a vertically-oriented mass forcing dipole, with in-built lifetime, and the ability to be advected between grid boxes (on a 15 km horizontal grid). The scheme had many problems and was probably not implemented in an optimal way. Nevertheless, it was an early (and brave) attempt to confront the problem that now faces NWP modellers – that the ensemble assumption cannot be considered valid for convective motions within a single grid box. This does not imply that the ensemble concept has to be abandoned – only that the effects of convection needs to be spread across many grid boxes in a computationally-efficient manner.

The notion that convective parametrization represents the statistical effect of many clouds is fundamental to its validity. But there are classes of convective activity for which this assumption cannot hold (on the time and space scales involved in NWP). For example, mesoscale convective systems are effectively isolated events : one would wish to see some explicit representation of these in NWP models. The current performance of limited-area models in regard to mesoscale convective system effects is of some interest and a case study project to study this has been initiated at the UK Met. Office. If it were possible to relate the likely occurrence of MCS systems to the distribution of CAPE and the synoptic/topographic forcing in an NWP model, then the imposition of a distributed mass forcing function may provide a suitable representation.

References

Cotton, W.R., M-S Lin, L. McAnelly and C.J. Tremback (1989) A composite model of mesoscale convective complexes. *Mon. Weather Rev.*, **117**, 765–783.

Davis, C. A. and Weisman, M. L. (1994) Balanced dynamics of mesoscale vortices produced in simulated convective systems. *J. Atmos. Sci.*, **51**, 2005–2030.

Gage, K. S. (1979) Evidence for a $k^{-5/3}$ law inertial range in mesoscale two-dimensional turbulence. *J. Atmos. Sci.*, **36**, 1950–1954.

Gage, K. S. and Nastrom, G. D. (1986) Theoretical interpretation of atmospheric wavenumber spectra of wind and temperature observed by commercial aircraft during GASP.*J. Atmos. Sci.*, **43**, 729–739.

Gray, M.E.B. (1996) Geostrophic adjustment following deep convection. Ph.D Thesis. University of Reading.

Gray, M.E.B., G.J. Shutts and G.C. Craig (1996) The role of mass transfer in describing the dynamics of mesoscale convective systems (submitted to Quart. J. Royal Met. Soc.)

Green, J.S.A. , F.H. Ludlam and J.F.R. McIlveen (1966) Isentropic relative-flow analysis and the parcel theory. *Quart. J. Roy. Meteor. Soc.*, **92**, 210–219.

Haynes, P.H. and McIntyre, M.E. (1987) On the evolution of vorticity and potential vorticity in the presence of diabatic heating and frictional or other forces. *J. Atmos. Sci.*, **44**, 828–841.

Haynes, P.H. and McIntyre, M.E. (1990) On the conservation and impermeability theorems for potential vorticity. *J. Atmos. Sci.*, **47**, 2021–2031.

Hedstrom, K. and L. Armi (1988) An experimental study of homogeneous lenses in a stratified rotating fluid. *J. Fluid Mech.*, **191**, 535–556.

Lilly, D. K. 1983. Stratified turbulence and the mesoscale variability of the atmosphere. *J. Atmos. Sci.*, **40**, 749–761.

Maddox, R.A. (1983) Large-scale meteorological conditions associated with midlatitude, mesoscale convective complexes. *Mon. Weather Rev.*, **111**, 1475–1493.

Moncrieff, M.W. and J.S.A. Green (1972) The propagation and transfer properties of steady convective overturning in shear. *Quart. J. Roy. Meteor. Soc.*, **98**, 336–352.

Moncrieff, M.W. (1992) Organized convective systems : Archetypal dynamical models, mass and momentum flux theory, and parametrization. *Quart. J. Roy. Meteor. Soc.*, **118**, 819–850.

Raymond, D.J. and Jiang, H. (1990) A theory for long-lived mesoscale convective systems. *J. Atmos. Sci.*, **47**, 3067–3077.

Shutts, G.J. (1994) The adjustment of a rotating, stratified fluid to localized sources of mass. *Quart. J. Roy. Meteor. Soc.*, **120**, 361–386.

Shutts, G.J. (1987) Balanced flow states resulting from penetrative slantwise convection. *J. Atmos. Sci.*, **44**, 3363–3376.

Shutts, G.J., M. Booth and J. Norbury (1988) A geometric model of balanced, axisymmetric flows with embedded penetrative convection. *J. Atmos. Sci.*, **45**, 2609–2621.

Shutts, G.J. (1995) An analytical model of the balanced flow created by localized convective mass transfer in a rotating fluid. *Dyn. Atmos. Ocean*, **22**, 1-17.

Shutts, G.J. and M.E.B. Gray (1994) A numerical modelling study of the geostrophic adjustment process following deep convection. *Quart. J. Roy. Meteor. Soc.*, **120**, 1145–1178.

Vallis, G.K., G.J. Shutts and M.E.B. Gray (1996) Balanced mesoscale motion and stratified turbulence forced by convection. (submitted to Quart. J. Royal Met. Soc.)