RECENT ESTIMATES OF DIABATIC FORCING ON THE PLANETARY SCALE: A REVIEW

Eero Holopainen
Department of Meteorology, University of Helsinki
00100 HELSINKI, Finland

Summary: A review of recent diagnostic estimates of the diabatic heating and related quantities is presented, and the uncertainties involved discussed. Some new results are presented based primarily on the ECMWF reanalyses of the FGGE data.

1. INTRODUCTION

Differential heating of the air between different latitudes is known to be the "prime mover" of the general circulation of the atmosphere. On the average, the forcing associated with this heating has to be balanced by damping caused by friction and diabatic processes that operate mainly on spatial scales much smaller than those at which the forcing occurs.

Our knowledge of the forcing and damping processes in the atmosphere is unsatisfactory. For example, an unsolved problem in the area of atmospheric energetics is the role of the different components of diabatic heating (radiation, release of latent heat, turbulent heat flux divergence) in the budget of available potential energy of the large-scale eddies. Another fundamental problem concerns the dissipation of kinetic energy in the free atmosphere: comparable estimates of the globally-averaged dissipation rate in the upper troposphere by Kung and Tanaka (1983) differ by a factor of five!

The generation of the available potential energy by the net diabatic heating as well as the dissipation of kinetic energy are, in studies of atmospheric energetics, normally evaluated as residuals. The results are sensitive to errors in the estimates of vertical velocity, which has been notoriously difficult to estimate from observations. New data and, in particular, developments in data assimilation methods during the last two decades have changed the situation considerably.

Diabatic forcing is in this report understood as comprising all nonadiabatic processes. In this broad sense forcing includes not only those processes that maintain circulation but also those that damp it (i.e. exert negative forcing).

Section 2 contains a few notes on the methodology of estimating the global distribution of diabatic heating from circulation data, and reviews some recent estimates obtained as the residual in the thermodynamic energy equation after using circulation data to evaluate all the other terms. Section 3 discusses the usefulness of momentum residuals obtained in an analogous way. Some recent estimates of the generation of available potential energy and dissipation of the kinetic energy in the atmosphere are reviewed in Section 4. Section 5 discusses residual estimates of the damping of potential enstrophy in the large-scale transient eddies in the atmosphere.

Most of the figures presented in this report show results from work carried out at the Department of Meteorology, University of Helsinki using global data for February 1979, the main full month of the Special Observing Period I of the Global Weather Experiment or FGGE. The data used consist of the first (referred to in the following as "MAIN") level IIIb analyses prepared by the ECMWF (European Centre for Medium Range Weather Forecasts), GLA (Goddard Laboratory for Atmospheres) and GFDL (Geophysical Fluid Dynamics Laboratory), and of the initialized reanalyses (referred to as "FINAL") by the ECMWF (Uppala, 1986). Emphasis in the paper is on the results derived from the ECMWF data.

2. ESTIMATING THE ATMOSPHERIC HEAT SOURCES AND SINKS FROM CIRCULATION DATA

2.1 Method

The discussion in this subsection is based on Holopainen and Fortelius (1986).

The diabatic heating per unit mass, Q, can be written formally as

$$Q = Q_R + Q_C + Q_T + Q_{diss} \qquad (1)$$

where Q_R is the net radiational heating, Q_C heating due to condensation processes, Q_T heating due to subgrid scale turbulent fluxes, and Q_{diss} heating due to dissipation of kinetic energy. Because Q_{diss} is much smaller than the other heating terms, one can write

$$Q \approx Q_R + Q_C + Q_T \ . \tag{2}$$

A direct way of estimating Q is to estimate the terms Q_R, Q_C and Q_T from suitable observations. If the heating of the whole mass of an air column is considered, Q_R is given by the difference of the net radiation between the top and bottom of the atmosphere, Q_C is proportional to the intensity of precipitation measured at the surface and Q_T is given by the turbulent heat flux at the surface. Global distribution of these terms for long-term mean conditions was presented by Budyko (1963), whose "heat atlas" has been the basic reference in physical climatology for a quarter of a century. However, on a daily or even monthly basis the diabatic heating cannot yet be estimated with sufficient accuracy in this way in the global domain. This will become more feasible when space-based radars eventually make it possible to monitor the global distribution of precipitation and, hence, of Q_C, which is the most variable term in (2).

An indirect method of determining Q is provided by the thermodynamic energy equation which, in the pressure coordinate system, can be written as

$$Q = c_p(p/p_0)^{\varkappa}\left(\frac{\partial\theta}{\partial t} + \mathbf{V}\cdot\nabla\theta + \omega\frac{\partial\theta}{\partial p}\right). \tag{3}$$

Here, as usual, θ is potential temperature, $\mathbf{V}$ horizontal velocity, $\omega = dp/dt$, c_p is the specific heat of air at constant pressure, $\varkappa = R/c_p$ (R is the gas constant), and p_0=1000 hPa.

If series of daily three-dimensional fields of T,$\mathbf{V}$ and ω are available, all terms on the rhs of (3) can be estimated and thus an indirect "residual" estimate for the net diabatic heating obtained. The estimation of the diabatic heating from (3) is, in principle, straightforward. In practice, however, getting meaningful estimates is not so easy because of the large sensitivity of the residual term to errors in the circulation data. The largest sensitivity is to errors in the divergent wind and the associated vertical velocity or, stated otherwise, to the degree with which the horizontal wind and vertical velocity estimates in the data set used depict the true mass balance in the atmosphere.

Even if great care is taken to reduce errors of various kinds, the instantaneous grid-point values of the estimated diabatic heating contain much noise. Filtering or averaging results with respect to time (and space) must be resorted to. It is then meaningful to cast (3) into such a form that circulation statistics can be used directly. If a time-mean is taken of (3) such that the local rate of change of temperature becomes small, and the resulting equation is integrated over the whole mass of an air column, the following equation is obtained:

$$\{\bar{Q}\} = c_p\{\bar{\mathbf{V}}\cdot\nabla\bar{T}\} + c_p\left\{(p/p_0)^{\kappa}\bar{\omega}\frac{\partial\bar{\theta}}{\partial p}\right\} + c_p\{\nabla\cdot\overline{T'\mathbf{V}'}\} + c_p\left\{(p/p_0)^{\kappa}\frac{\partial}{\partial p}(\overline{\theta'\omega'})\right\}. \qquad (4)$$

As usual, a bar denotes the time average and a prime deviation from the time-average. Symbol { } denotes a vertical integral.

Equations similar to (4) can be also written for the water vapour. In these equations one can, if one so wishes, use the flux divergence form instead of the advective form not only for the eddy terms but also for the mean-flow terms.

2.2 A review of recent estimates of diabatic heating

Examples of atmospheric heating studies based on the use of circulation data are given in Table 1. There are basically two approaches. One may either use the basic observations at the rawinsonde stations (approach A) or large-scale daily analyses of the circulation variables (approach B).

The most recent study using approach A is by Savijärvi (1988), who reports on an attempt to extend the calculations by Carissimo et al. (1985) (dealing with the zonally-averaged

Table 1 Examples of studies of diabatic heating based on the use of circulation data

A Use of statistics based on radiosonde station data	B Use of statistics based on daily analyses	
	One set of analyses	Intercomparisons
Newell et al.(1974)	Lau (1979)	Bretherton et al. (1982)
Oort & Vonder Haar (1976)	Johnson et al.(1985)	Lorenc & Swinbank (1984).
Carissimo et al.(1985)	Hoskins (1985)	Boer (1986)
Savijärvi (1988)	Kasahara & Mizzi (1985)	Holopainen & Fortelius(1986)
	Chen & Baker(1986)	Masuda (1988)

conditions) to a full geographical domain. A recent application of approach B is by Masuda (1988), who calculates the zonally averaged energy fluxes and the associated heating for the full FGGE year from the "MAIN" analyses by ECMWF and GFDL.

The error bars in the heating estimates pose an important and difficult problem. Some feeling for the uncertainties involved may be obtained by comparing estimates from several sets of analyses which all refer to the same period. Differences in the heating estimates arise in this case due to the different usage of the basic observations and differences in the data assimilation methods.

In this area, the first intercomparison of heating estimates was made by Lorenc and Swinbank (1984), who compared the atmospheric energy flux divergence estimated from operational ECMWF and UK Meteorological Office analyses for July 1979. They gave a rather optimistic view of the accuracies involved. A more pessimistic result was obtained by Boer (1985), who compared energy fluxes and the associated heating obtained for July 1979 from FGGE level IIIb "MAIN" analyses by ECMWF and GFDL. The results by Holopainen and Fortelius (1986), who compared the heating estimates obtained from the ECMWF and GLA "MAIN" analyses for February 1979, represent an intermediate view.

Fig. 1a shows the global distribution of the vertically integrated diabatic heating in February 1979, as obtained from the "MAIN" analyses by ECMWF. Uninitialized wind analyses were used to calculate ω by using the continuity equation and imposing the mass balance requirement (for details see Holopainen and Fortelius, 1986). Fig. 1b shows the same pattern obtained from "FINAL" analyses by directly applying the initialized fields of V, T and ω in (4).

Figures 1a and 1b look rather similar. Features common to both panels are, for example, large positive values over the equatorial western Pacific, Brazil and and the "storm track" regions off the eastern coasts of North America and Eurasia.

Fig. 1c shows the difference ("FINAL"-"MAIN") in the heating. Ideally, this difference should be zero everywhere. Relatively large values are seen in Fig. 1c, however. These

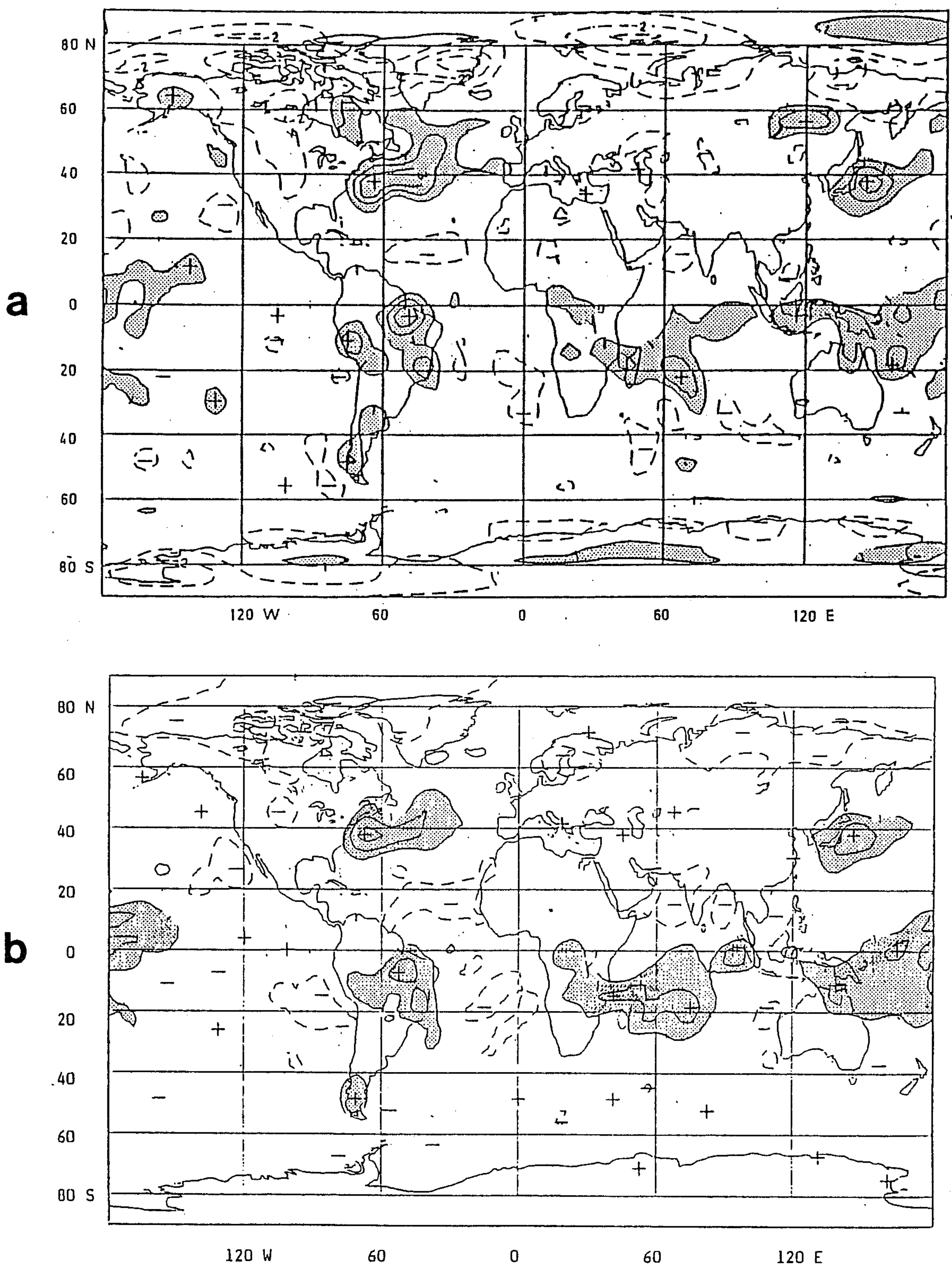


Fig. 1

(a) The vertically integrated net diabatic heating as determined from the ECMWF level IIIb "MAIN" circulation data by using expression (4). Units: 100 W m^{-2}. Contour interval is 100 W m^{-2}, but for the sake of clarity, the zero line is not shown. Areas with heating larger than 100 W m^{-2} are shaded. (From Holopainen and Fortelius, 1986).

(b) As in (a) but from the ECMWF level IIIb "FINAL" data.

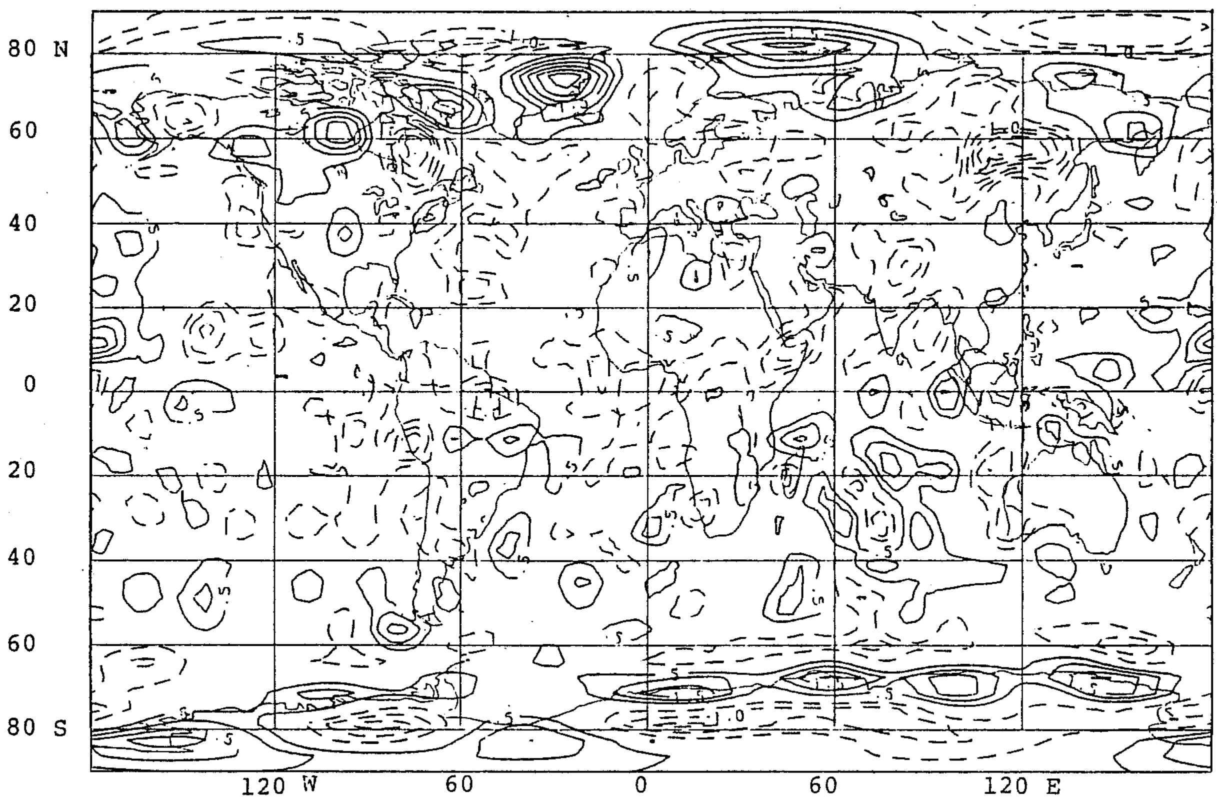


Fig. 1c

Difference ("FINAL"-"MAIN") between the heating distributions seen in Figs. 1b and 1a. Contour interval is 50 W m^{-2}.

are due to differences in the basic (level IIb) data, in the data assimilation system used to generate the two sets of analyses, and in the initialization. (In the "MAIN" set from ECMWF, the horizontal wind and geopotential fields are uninitialized.)

One can argue that many of the features seen in Fig. 1c are probably mainly due to errors in the values seen in Fig. 1a. As an example one can take the large heating values in Fig. 1a in eastern Siberia between 55° and 60°N around longitude 120° E. The only way large positive values of Q could arise in this region in February, where Q_R is likely to have a large negative and Q_T a small negative value, is to have (see (2)) large release of latent heat (Q_C>>0). However, the observed monthly precipitation totals (FBU, 1979) in this region for February 1979 were of the order of 10 mm, which corresponds to $Q_C \approx 10$ W m^{-2}. Thus, the total heating Q in this region should be negative as in Fig.1b.

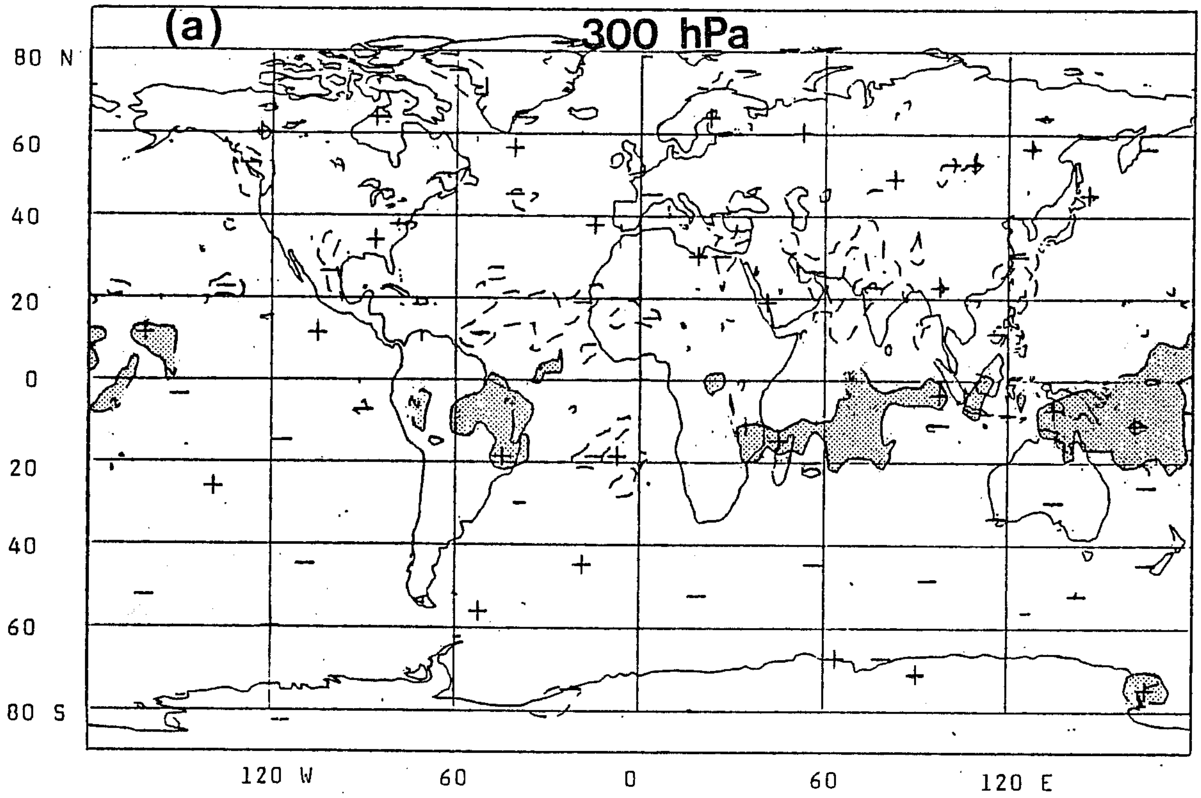
When averages are taken over the "Cage" areas (Bretherton et al, 1982), the ("FINAL"-"MAIN") heating differences from Fig. 1c are -27, +10, 0 and -20 W m^{-2} for "North Atlantic", "North Pacific", "North America" and "Eurasia" , respectively. The mean absolute value of these differences is thus 14 W m^{-2}. This is of the same magnitude as the value (9 W m^{-2}) reported by Holopainen and Fortelius (1986) for the absolute mean difference between the heating estimates derived from the ECMWF and GLA "MAIN" analyses.

2.3 Vertical distribution of heating

Fig. 2 shows, for the 700 hPa and 300 hPa levels, the heating pattern obtained from "FINAL" analyses. At the 700 hPa level (Fig. 2b) the "storm tracks" in the northern extratropics are clearly visible as local maxima of Q. These maxima are likely to result from large releases of latent heat and fluxes of sensible heat from the ocean surface.

The heating over the "storm tracks" is largely confined to the lower and middle troposphere: there are no indications of them at 300 hPa (Fig. 2a). In the tropics, however, the heating over the main convection areas appears to be much stronger at 300 hPa than at 700 hPa. This difference in the vertical distribution of heating between the tropics and the extratropics is clearly seen in Fig. 3, which shows the meridional cross section of the heating in February 1979. Qualitatively, the pattern appears realistic and agrees with that presented by Newell et al. (1972/1974).

Fig. 4 shows the vertical distribution of the the globally averaged diabatic heating. Positive values are seen in the lower troposphere and negative values in the upper troposphere. As discussed by Holopainen (1968) and Holopainen and Fortelius (1986), there is a natural global constraint that produces this kind of heating distribution. It can be formally derived by writing (3) in the flux divergence form and taking a time mean and a global average over an isobaric surface. One obtains



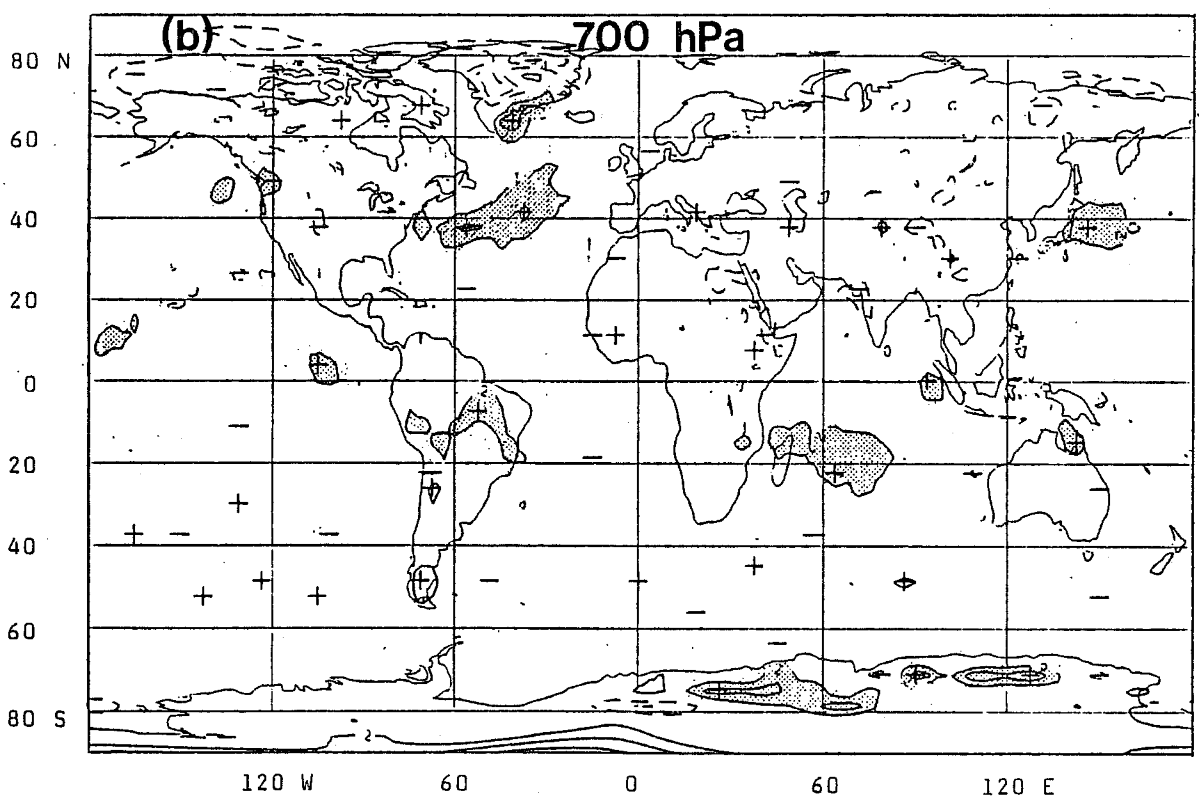


Fig. 2
The mean diabatic heating Q (in K/d) at (a) 300 hPa and (b) 700 hPa in February 1979, as evaluated from the ECMWF level IIIb "FINAL" data. Areas with heating larger than 2 K/d are shaded.

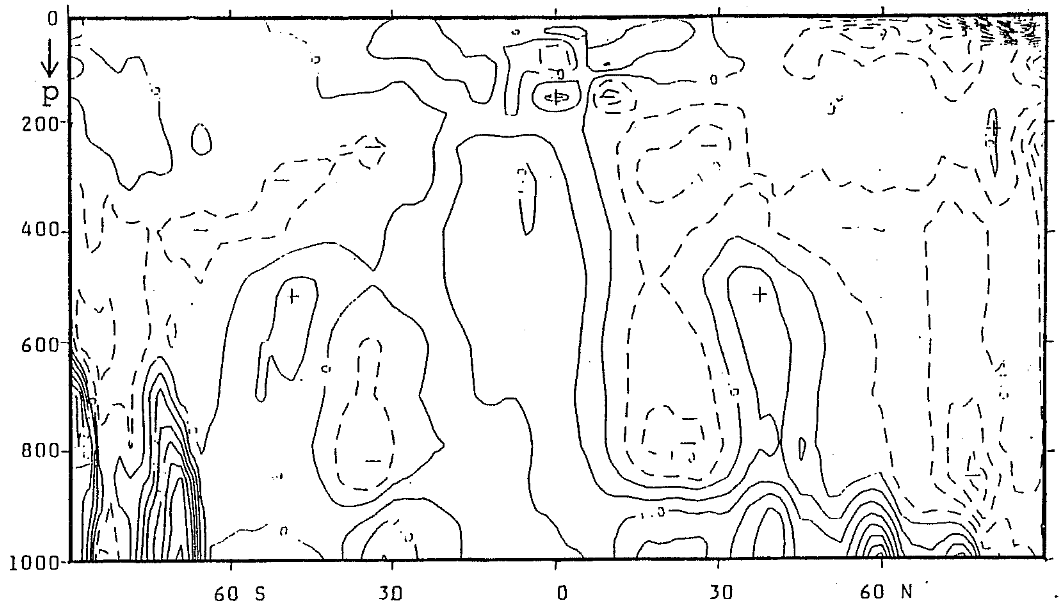


Fig. 3
Zonally-averaged time-mean diabatic heating Q (in K/d) in February 1979, as estimated from the ECMWF level IIb "FINAL" data.

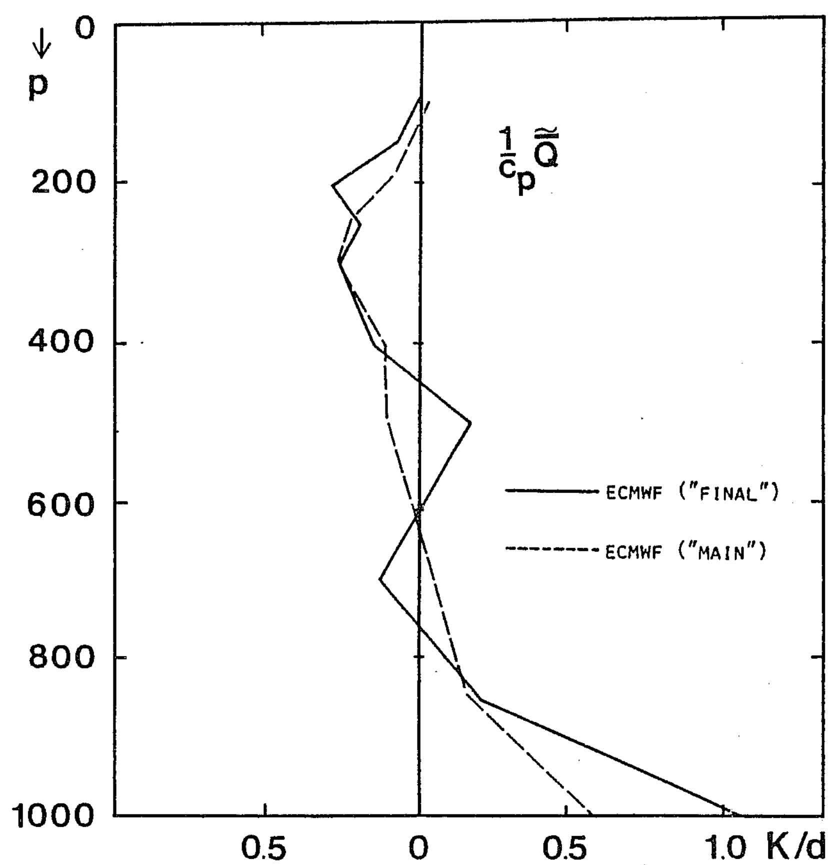


Fig. 4
As in Fig. 3 but averaged isobarically over the whole globe. Continuous (dashed) line shows values obtained from the "FINAL" ("MAIN") data set of the ECMWF.

$$\partial/\partial p(\widetilde{\theta\omega}) = \frac{1}{c_p}\left(\frac{p_o}{p}\right)^{\kappa} \tilde{Q} \tag{4}$$

(In (4), $\tilde{(\ )}$ denotes an area average on an isobaric surface.) The circulation must transport heat towards lower pressure for kinetic energy to be generated ($\widetilde{\theta\omega} < 0$). Furthermore, $\widetilde{\theta\omega}$ = 0 at the top and at the bottom of the atmosphere. Therefore, the l.h.s. of (4) and, accordingly, Q also has to be positive in the lower troposphere and negative in the upper troposphere.

As Fig. 4 shows, both "MAIN" and "FINAL" analyses by ECMWF give roughly the same vertical heating distribution. A major difference between the two data sets occurs at 700 hPa, where the FINAL analyses give cooling while "MAIN" analyses indicate warming; the opposite is true at 500 hPa. A closer analysis of the heating fields reveals that these differences in Fig. 4 arise mainly from the contribution of southern hemisphere. The cooling maximum in Fig. 3 in the 700-850 hPa layer at about 30°S occurs at much higher levels in the pattern derived from "MAIN" analyses (not shown), whereas the heating at around 500 hPa at 50° S in Fig. 3 is absent in the figure derived from the "MAIN" analyses.

The heating differences between the "MAIN" AND "FINAL" are most likely connected with the differences in the vertical modes of the divergence in the two data sets.

2.4 "Residual" heating versus parameterized heating

In the forecast models which provide the first-guess field for the analyses of circulation variables, the heating components Q_R, Q_C and Q_T are parameterized in some way. Ideally, the parameterized total heating should be equal to the residual estimate one obtains from the circulation data by using (3).

Figs. 5a and 5b show the vertically-integrated residual heating and the parameterized model heating, respectively, as determined from the GLA data set. (The parameterized heating from the ECMWF assimilation model was not available by the a

b

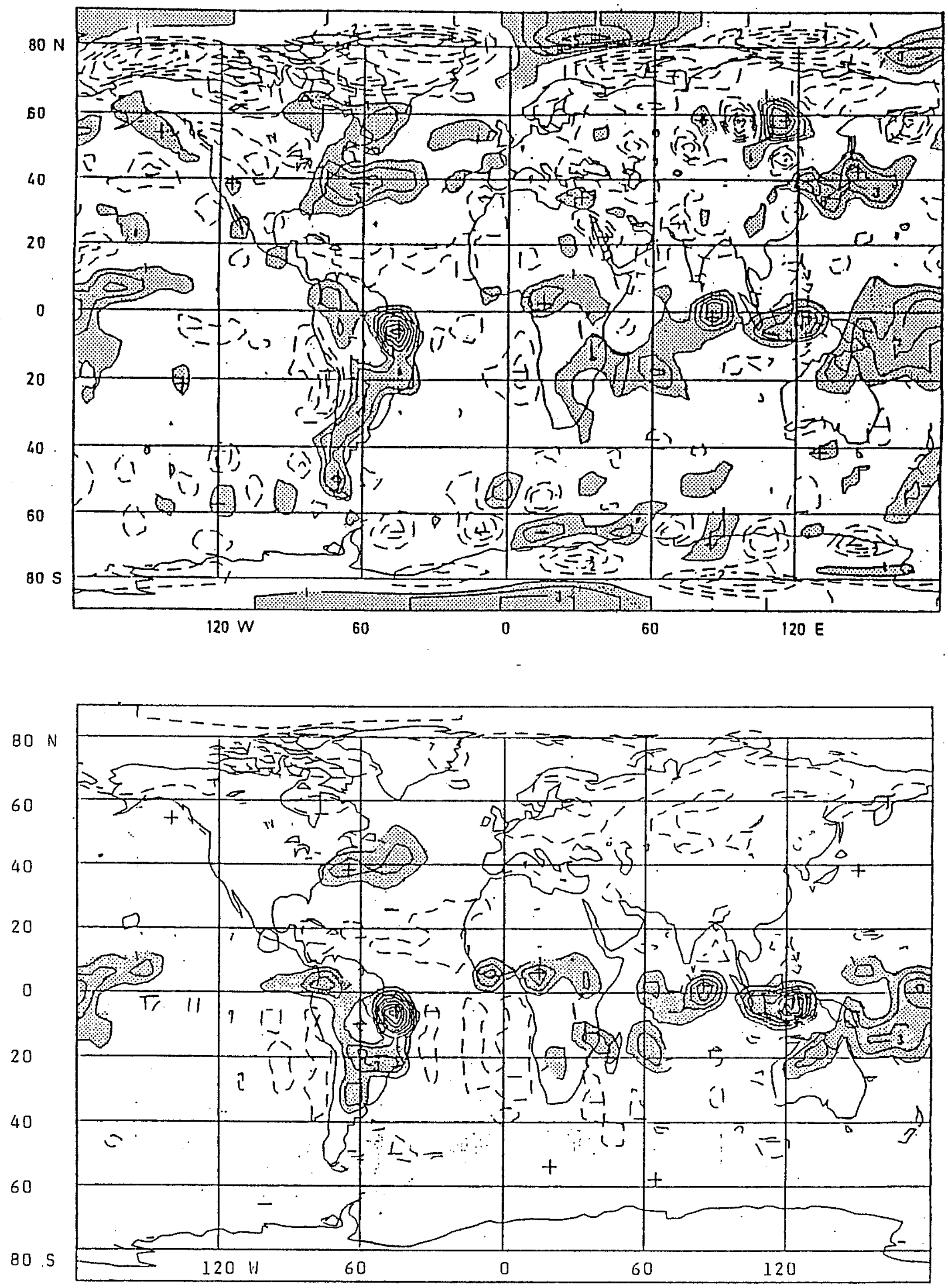


Fig. 5

(a) As Fig. 1a but from the GLA level IIIb circulation data for February 1979. (Holopainen and Fortelius, 1986).

(b) The vertically-integrated diabatic heating in February 1979 as parametrized in the GLA forecast model, which produced the first-guess field to the GLA level IIIb circulation data.

time of writing this report). In the tropics there appears to be a good similarity between the two fields in Fig. 5. This similarity should not be interpreted as indicating that we know (and even that we can well parametrize) the heating in the tropics. Probably a more realistic interpretation is that in the tropics, where the diabatic heating and vertical velocity are intimately coupled, the patterns of divergent wind/vertical velocity/heating in the GLA analyses are determined mainly by the way in which the heating is parameterized in the forecast model. The same is probably true in other data assimilation systems as well. Clearly, estimates of the diabatic heating in the tropics by the residual technique are at present less reliable than those in the extratropics.

3. MOMENTUM RESIDUALS

Section 2 showed that, for the time being, the distribution of the net diabatic heating can, at least in the extratropics, be best estimated as a residual term from the thermodynamic energy equation. One can then legitimately ask whether the corresponding residual term in the equations of motion ("friction") gives information about the effect of subgrid-scale processes on the resolved large-scale flow. Can one, for example, extract from the large-scale momentum budget any evidence for the effect of the gravity wave drag, so much discussed (e.g. Palmer et al. (1986) in recent years? So far, the main observational evidence for the existence of such a drag comes from the residual term in the longitudinally averaged budget of zonal momentum (Swinbank, 1985). Studies of the local momentum budget (e.g. Holopainen et al., 1980) have yielded in this respect inconclusive results, partly because of the lack of data on vertical velocity. Are local values of the momentum residuals from the FGGE analyses more useful in this respect?

The time-mean budget of zonal momentum at an arbitrary point can be written as

$$0 = A^u + f\bar{v}_a + R^u \qquad (5)$$

where A^u is the sum of the three-dimensional momentum flux convergence and the metric term, f is the Coriolis parameter, and v_a the ageostrophic meridional wind component. R^u is the zonal component of the residual force, which represents the effects of subgrid-scale motions.

Fig. 6 shows the zonally-averaged zonal residual force R^u obtained from the ECMWF "FINAL" analyses. The lower troposphere features are qualitatively what one might expect: positive values in the region of easterlies, negative values in the region of westerlies. (Due to interpolations across mountains, the low troposphere values in Fig. 6 should not be considered quantitatively).

In the free atmosphere the only conspicuous feature in Fig. 6 are the negative values of R^u (up to -4 m s^{-1}/d) in the region of the subtropical jet stream around 30°N. The geographical distribution of R^u (not shown) reveals that this feature arises not so much from the mountainous regions as from the region of the Pacific Ocean. Although there are a lot of single level data for this oceanic region during FGGE

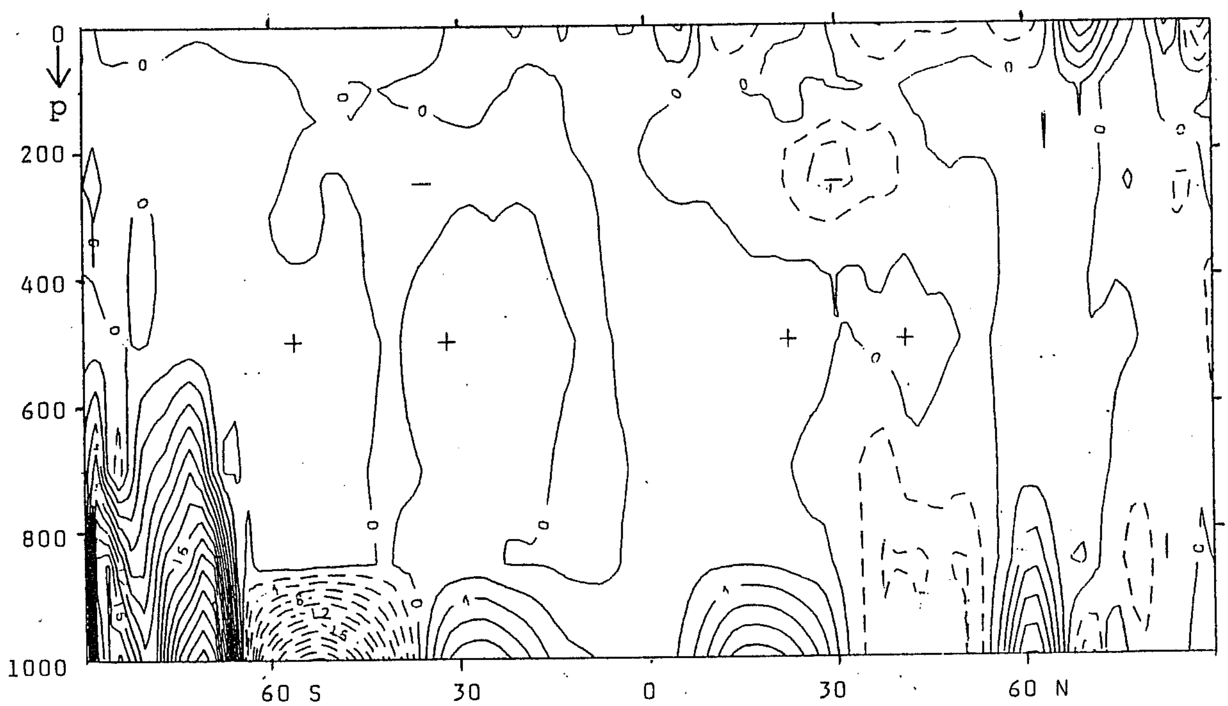


Fig. 6
Zonally-averaged mean residual zonal force (in 10^{-5} m s^{-2}) in February 1979, obtained from the ECMWF level IIIb "FINAL" data set.

(Hollingsworth et al., 1986), it is still likely that the analysis of the ageostrophic meridional velocity, which is an important ingredient in the zonal momentum budget, is in this region dictated primarily by the model's first-guess field. If this is true the large negative values of R^u in Fig. 6 may be just a reflection of the large parameterized friction in the model, and does not necessarily depict what happens in real nature. (The initialization may also influence the vertical distribution of v_a and R^u).

Fig. 7 shows the budget of zonal momentum in the free atmosphere over the North American continent, which is covered with a homogeneous network of radiosonde stations. It is seen that A^u (essentially the horizontal convergence of zonal momentum) is practically the same from "MAIN" and "FINAL" data.

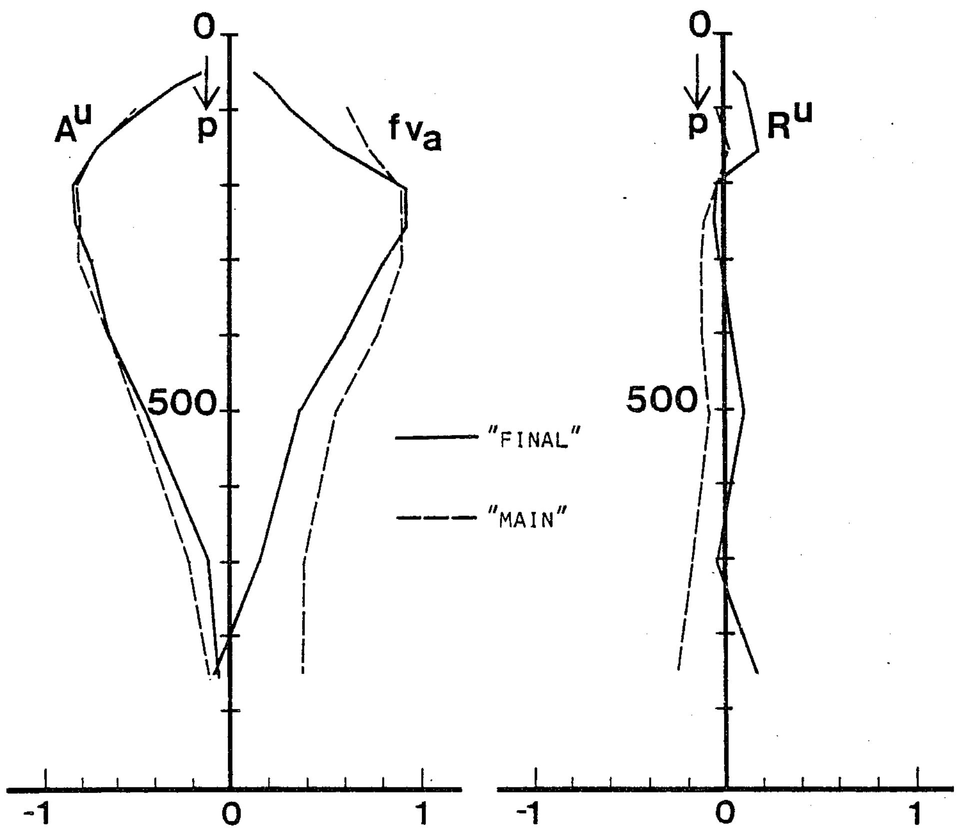


Fig. 7
Budget of zonal momentum (in 10^{-4} m s^{-2}) over North America in February 1979. The values obtained from the ECMWF initialized "FINAL" data are shown by a solid line, those from the ECMWF uninitialized "MAIN" data by a dashed line.

The term $f\bar{v}_a$, however, is very different depending on the data set used, as consequently is R^u also. The residual term calculated from the initialized "FINAL" analyses has a smaller magnitude than that obtained from the "MAIN" analyses.

It can be argued that only the rotational part of the residual force is worth considering. The global field of the residual force vector $(R^u\mathbf{i}+R^v\mathbf{j})$ from the "FINAL" data set was partitioned into divergent and rotational components. The zonal component of the nondivergent residual force vector, averaged over North America, turned out to be practically equal to the "FINAL" R^u values seen in Fig. 7.

The momentum residual R^u obtained for North America from the GLA data set for February 1979 (not shown) turned out to have a larger magnitude than in the case of either of the two ECMWF data sets. R^u from the GFDL data was positive at all levels and had the largest magnitude.

It would appear that an interpretation in physical terms of the momentum residuals obtained by using the fields produced by present-day data assimilation systems is hardly justified.

4. SOME ASPECTS OF THE ENERGETICS OF THE GENERAL CIRCULATION

The most comprehensive recent study of the atmospheric energetics is that by Arpe et al. (1986). They calculated all the energy conversion terms from the ECMWF initialized analyses and 12-hour forecasts. The dissipation of kinetic energy and the generation of the available potential energy were then obtained as residuals.

Fig. 8 shows the vertical profile of the residual estimate of the globally-averaged dissipation of eddy kinetic energy by Arpe et al.(1986), based on the 12-hour forecasts. Besides a maximum in the boundary layer it shows a second maximum in the upper troposphere, in qualitative agreement with the findings of Kung and his collaborators (e.g. Kung and Baker, 1975).

Arpe et al. warn against taking their dissipation estimates too literally: "We do not know how far these calculations are influenced by the analysis-forecast system or by observations....". In fact the dissipation evaluated from the

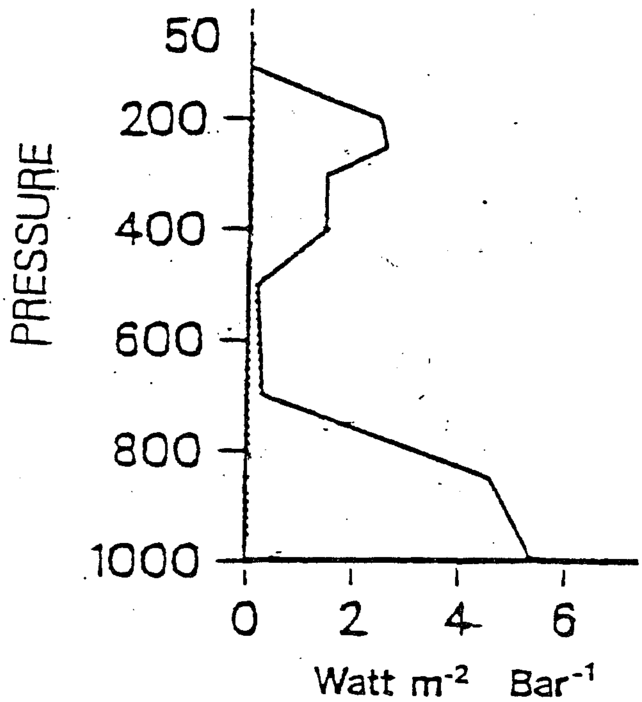


Fig. 8
Vertical profile of the globally averaged rate of dissipation of eddy kinetic energy. Unit: W m^{-2}/bar. (From Arpe et al., 1986)

forecast fields by the method of Arpe et al. should, if properly evaluated, be equal to the dissipation one obtains from the parameterization formulas used for the frictional processes in the model. The present parameterizations may, however, not be realistic and, accordingly, the dissipation values seen in Fig. 8 may not be good estimates of the real dissipation.

The dissipation estimates by Kung et al. for North America may not be representative of hemispheric or global conditions (see e.g. Holopainen and Eerola, 1979), and the real ("observed") dissipation values comparable to those in Fig. 9 are not known. This is true not only for the free atmosphere, where the largest uncertainty at present probably lies, but to a smaller extent also for the planetary boundary layer. Our ignorance concerning the frictional processes and the concomitant deficiencies in their parametrization is possibly one of the reasons for the systematic errors in the large-scale models today.

Fig. 9 shows the mean annual course of the generation of eddy available potential energy, estimated by Arpe et al. as a residual term in the budget of eddy available potential energy. The annual course of this term (larger values in summer than in winter) appears qualitatively realistic and agrees with that presented by Oort and Peixoto (1983). Quantitatively, however, large uncertainties remain. For example, the annual mean value for the whole globe is -0.01 W m^{-2} from the initialized analyses and +0.38 W m^{-2} from the 12 hour forecast fields.

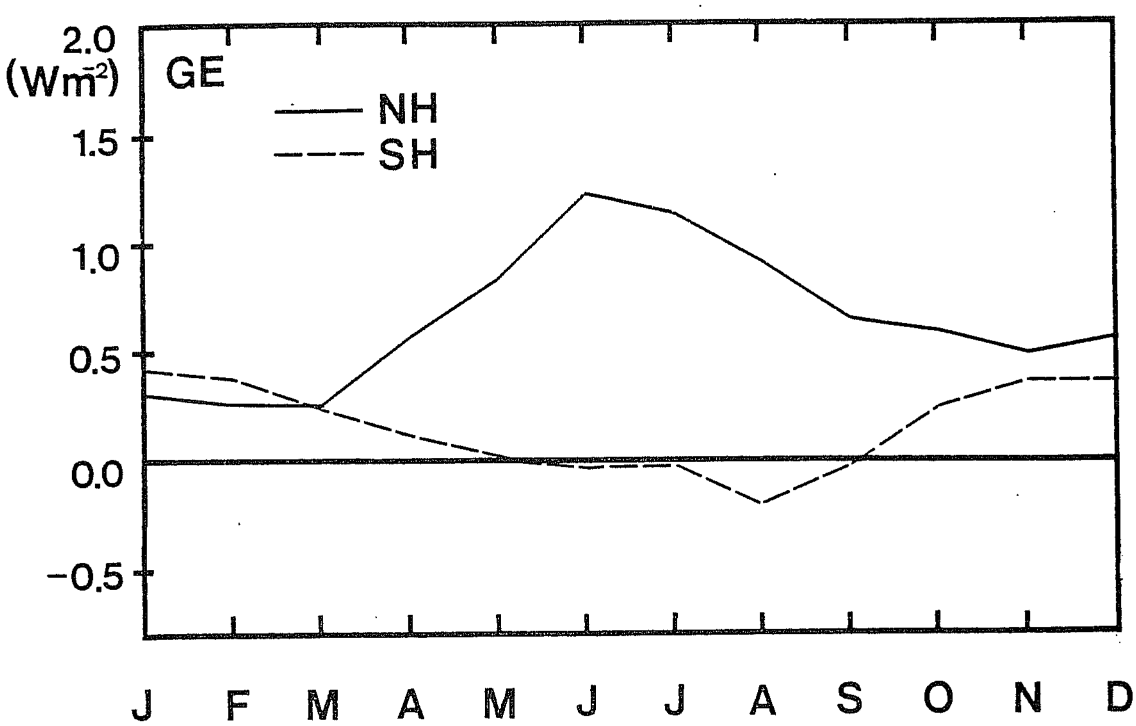


Fig. 9
Annual cycle of the generation of the eddy available potential energy (GE) for the northern (solid line) and southern (dashed line) hemisphere. Unit: W m^{-2}. (From Arpe et al., 1986).

5. SOME ASPECTS OF THE POTENTIAL ENSTROPHY BUDGET OF THE GENERAL CIRCULATION

The conservation law for potential vorticity is the backbone of dynamic meteorology. Documentation of atmospheric large-scale behaviour in terms of potential enstrophy should, accordingly, be at least as important as in terms of energy. Some work based on a quasi-geostrophic framework was recently reported by Holopainen and Fortelius (1987).

A description of the atmospheric general circulation in terms of potential enstrophy (Fig.10a) is qualitatively the same as in terms of energy (Fig.10b) : the time-mean flow that is generated by latitudinal differences in diabatic heating breaks down into transient eddies, which in turn are damped by diabatic and frictional processes.

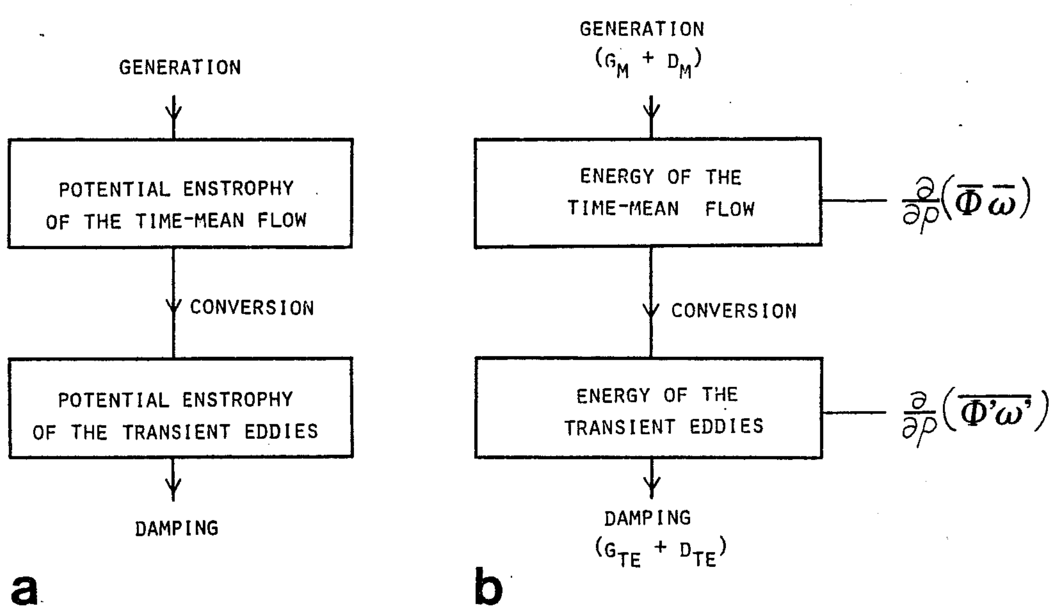


Fig. 10
Schematic diagram for the general circulation of the atmosphere in terms of
(a) potential enstrophy.
(b) energetics (G_M denotes the generation of the available potential energy and D_M the dissipation of kinetic energy of the time-mean flow; G_{TE} and D_{TE} are the corresponding quantities for the transient eddies. As usual, Φ denotes geopotential.

Potential enstrophy considerations have at least two formal advantages over those of energetics. First, estimates of terms in the potential enstrophy budget involve only the relatively accurately-known fields of temperature and horizontal wind, whereas energetics depend also on the poorly-known fields of vertical velocity (and horizontal wind divergence). Secondly, potential enstrophy calculations provide information for individual pressure levels, whereas estimates of the corresponding energetic processes can probably be worked out with comparable accuracy only for the vertically-integrated case. Thirdly, the potential enstrophy budget does not directly depend on the presence of mountains and the Ekman layer. Mountains have, however, a direct effect on the energy conversion from the mean flow to eddies, and this effect is difficult to estimate accurately.

In the steady state, the budget equation for the isobarically averaged budget of the quasi-geostrophic potential enstrophy of the transient eddies (TE) is (for details see Holopainen and Fortelius, 1987):

$$\partial/\partial t(\overline{\widetilde{\tfrac{1}{2}q'^2}}) = 0 = -\overline{\widetilde{\mathbf{V}'q'\cdot\nabla\bar{q}}} + \overline{\widetilde{D'q'}} \qquad (6)$$

In (6), q is the (quasi-geostrophic) potential vorticity and D the source of q due to diabatic and frictional processes:

$$D = -f\ \partial/\partial p\left(\frac{(c_p)^{-1}(p_0/p)^{\kappa}\tilde{Q}''}{S}\right) - g\ \partial/\partial p(\mathbf{k}\cdot\mathrm{curl}\,\boldsymbol{\tau}) + \mathbf{k}\cdot\mathrm{curl}\ \mathbf{F}_H \qquad (7)$$

In (7) $S = -\partial\tilde{\bar{\theta}}/\partial p$, $\boldsymbol{\tau}$ is the turbulent stress due to vertical subgrid-scale momentum fluxes and $\mathbf{F}_H$ the horizontal force due to the corresponding horizontal fluxes. As usual, $\mathbf{k}$ is the unit vector in the vertical.

According to (7), the source/sink term D is due to stabilization of the air due to differential heating in the vertical, and due to horizontal and vertical diffusion of vorticity by subgrid-scale processes.

In (6) $\overline{D'q'}$ represents the effect of diabatic and frictional processes on the TE potential enstrophy. According to (6) and Fig. 10a this effect has, on the average, to be balanced by the conversion of potential enstrophy from the time-mean flow to the TE's. This conversion is represented by the term $-\widetilde{\overline{\mathbf{V}'q'}}\cdot\nabla\overline{q}$, which can be evaluated (and, hence, a residual estimate of $\widetilde{\overline{D'q'}}$ be obtained) having information on the fields of temperature and horizontal wind only.

Fig. 11 shows for February 1979 the northern hemisphere averages of $\widetilde{\overline{D'q'}}$, $\widetilde{\overline{\frac{1}{2}q'^2}}$ and the "damping coefficient"

$$\nu(p) = -\widetilde{\overline{D'q'}}/(\widetilde{\overline{\tfrac{1}{2}q'^2}}).$$

$\nu(p)$ can be considered as a gross measure of the rate of damping of TE potential enstrophy; $1/\nu$ is the corresponding time scale.

$\overline{D'q'}$ is seen to be negative at nearly all levels indicating damping of the TE potential enstrophy; the time scale of this damping is about 10 days. As discussed by Holopainen and Fortelius (1987), both frictional and diabatic processes are likely to be important in this damping.

The degree of sensitivity of the $\widetilde{\overline{D'q'}}$ estimates to differences in the data sets is revealed by Fig. 12a, which shows $\widetilde{\overline{D'q'}}$ as calculated from four FGGE level IIIb data sets for February 1979. Although all curves indicate damping ($\widetilde{\overline{D'q'}}<0$), a large scatter is noticable. This scatter appears to a large extent to be due to differences between the globally-averaged static stability in the three data sets. In the quasi-geostrophic framework, the average static stability is considered to be an "external" parameter, a function of pressure only. It is therefore meaningful to compare the estimates of $\widetilde{\overline{D'q'}}$ obtained when the same reference static stability is used in all data sets. Such a comparison is provided by Fig. 12b. It is seen that the scatter is considerably reduced compared to that in Fig. 12a. It is further seen that the ECMWF "MAIN", ECMWF "FINAL" and GLA data sets give

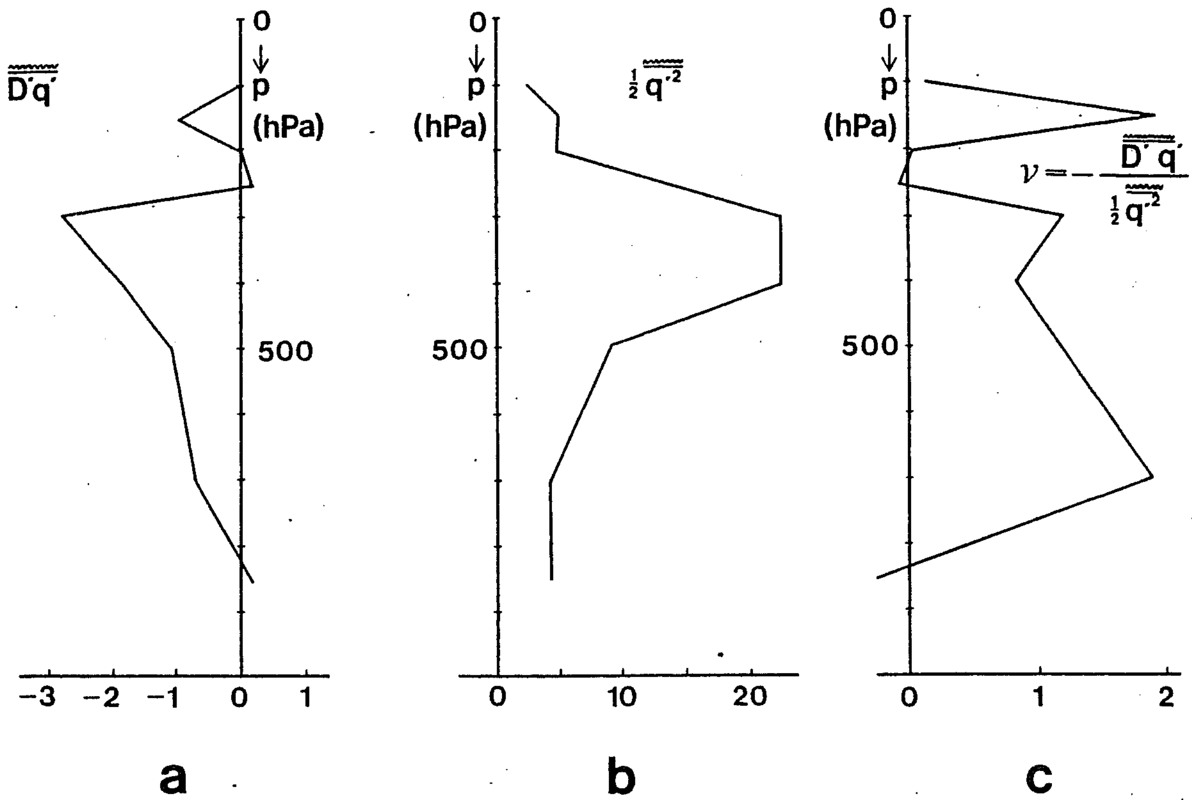


Fig. 11

Northern hemisphere averages for February 1979 of

(a) $\overline{\widetilde{D'q'}}$, the effect of diabatic and frictional processes on the transient eddy potential enstrophy (in 10^{-15} s^{-3})

(b) $\frac{1}{2}\overline{\widetilde{q'^2}}$, the TE potential enstrophy (in 10^{-10} s^{-2})

(c) the "damping coefficient" $\nu(p)=-\overline{\widetilde{D'q'}}/(\frac{1}{2}\overline{\widetilde{q'^2}})$ (in $10^{-6}s^{-1}$).

The estimates have been obtained by using the level IIIb "MAIN" analyses by the ECMWF. (From Holopainen and Fortelius, 1987).

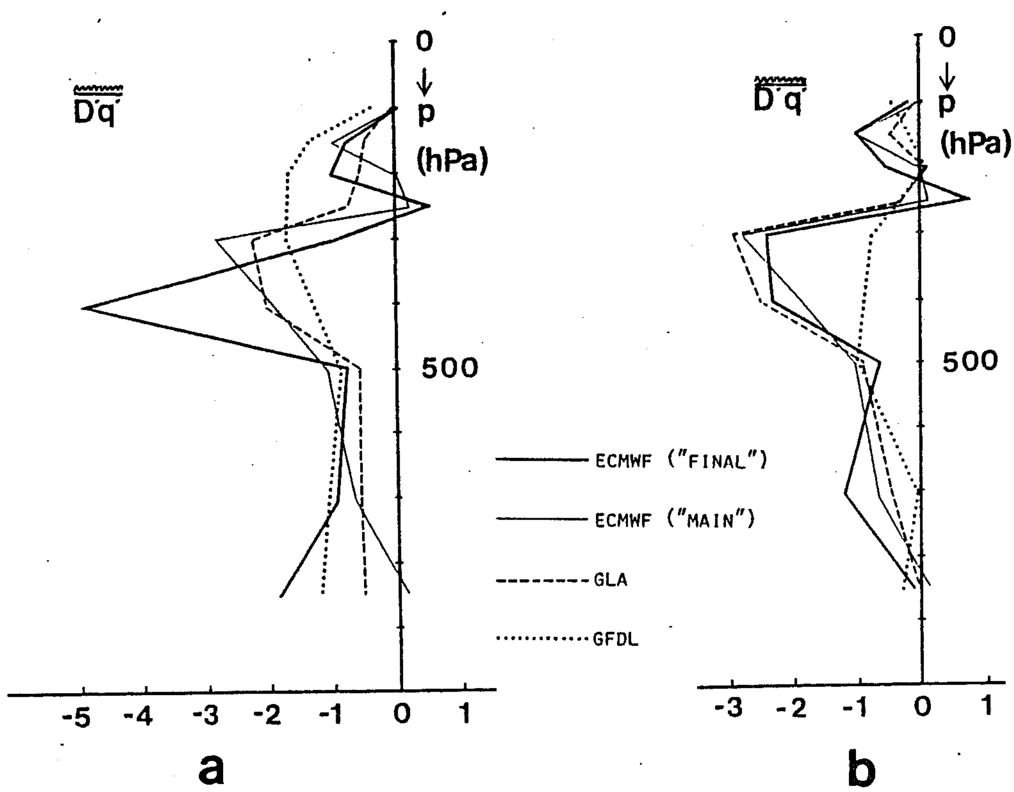


Fig. 12

Residual estimate of $\widetilde{\overline{D'q'}}$ (in 10^{-15} s^{-3}) over the northern hemisphere in February 1979, estimated from various circulation data

(a) results obtained when the global-mean static stability is calculated for each data set separately

(b) results obtained by using the same value (obtained from the ECMWF "MAIN" data set) for the global-mean static stability in all data sets.

damping profiles which are rather similar to each other. In the upper troposphere, where the damping appears to be largest, the GFDL data give much smaller values than the other data sets.

The profile of $\widetilde{\overline{D'q'}}$ is a fundamental characteristic of the general circulation. It could turn out to be a useful quantity in the validation and intercomparison of the atmospheric GCM's.

Instead of the quasi-geostrophic vorticity and pressure coordinates one should in possible later studies perhaps use "Ertel's potential vorticity" and the isentropic framework (e.g. Townsend and Johnson, 1985; Hoskins et al., 1985). The basic features of $\widetilde{\overline{D'q'}}$ will, however, hardly be different from those discussed above.

6. CONCLUSIONS AND REMARKS

Uncertainties in the continental scale averages of the vertically-integrated monthly-mean diabatic heating are at least of the order of 10 W m^{-2} in the northern extratropics and larger than this in the tropics and in the southern hemisphere.

Despite these large uncertainties in the heating estimates the situation is encouraging. Compared with the situation, say, ten years ago, meteorologists now appear to be able to describe the planetary-scale driving force of the general circulation much more accurately, at least on time scales of one month or longer. It remains to be seen whether useful estimates can be derived for shorter time scales.

What is clearly needed is independent data, against which the residual heating estimates derived from circulation data can be compared. An example of such independent data is the satellite observations of the net radiation at the top of the atmosphere. In principle these can be used in the validation of the heating estimates if the latent heat is included in the type of calculations reported here (Oort and Vonder Haar, 1976). Unfortunately, satellite radiation data have large uncertainties of their own (e.g. Jakobowitz and Tighe, 1984; Hartmann et al., 1986), and there is no certainty yet that they will be useful in this validation. In the near future, the greatest progress in our ability to describe the diabatic heating in the atmosphere will probably come, not from entirely new data, but rather from improvements in the data assimilation methods. On the other hand, any possible new source of data (as, for example, global precipitation measurements by satellite-borne radars) can eventually be included in the data assimilation system.

It would appear that the local residual terms in the budget of momentum and kinetic energy, obtained from large-scale analyses, are so much affected by the data assimilation system (forecast model, initialization etc.) that their interpretation in physical terms as "observed friction" or "observed dissipation" is at present probably not justified.

The "observed" dissipation of kinetic energy of the large-scale flow in the free atmosphere is still highly uncertain. In the globally-averaged case there must be some loss of kinetic energy in the upper troposphere from the large-scale flow to smaller scales (otherwise how could we have e.g. clear air turbulence?). But whether this average upper level dissipation is of the same order of magnitude or an order of magnitude smaller than the dissipation in the boundary layer, at present we simply do not know.

In contrast to the case of energetics, which are heavily dependent on the poorly-known divergent wind, the major features of the potential enstrophy budget are (at least in the extratropics) determined by the more accurately-known fields of temperature and rotational wind. Accordingly, potential enstrophy might turn out to be a useful tool in diagnosing the net effect of diabatic and frictional processes in the atmosphere, and in the verification and intercomparison of large-scale models.

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REFERENCES

Arpe, K., C. Brankovic, E. Oriol and P. Speth, 1986: Variability in time and space of energetics from a long series of atmospheric data produced by ECMWF. Contrib. Atm. Phys., 59, 321-355.

Boer, G.J., 1986: A comparison of mass and energy budgets from two FGGE data sets and a GCM. Mon. Wea. Rev., 114, 885-902.

Bretherton, F.B., D.M.Burridge, J. Creease, F.W. Dobson, E.B. Kraus and T.H. Vonder Haar, 1982: The "CAGE" experiment: A feasibility study. WCP-22. WMO Secretariat, Geneva, 95 pp.

Budyko, M.I. (Ed), 1963: Atlas of the heat balance of the Earth. Kartafabrika Gosgeoltehizdata, Leningrad, 69 pp. (In Russian.)

Carissimo, B.C., A.H. Oort and T.H. Vonder Haar, 1985: Estimating the meridional energy transport in the atmosphere and oceans. J. Phys. Oceanogr., 15, 82-91.

Chen, T-C. and W.E. Baker, 1986: Global diabatic heating during FGGE SOP-1 and SOP-2. Mon. Wea. Rev.,114, 2578-2589.

FUB, 1979: Beilage zur Berliner Wetterkarte des Instituts für Meteorologie der Freien Universität Berlin: Nordhemisphärische Wetterlage im Februar 1979.

Hartmann, D.L., V. Ramanathan, A. Berroir and G.E. Hunt, 1986: Earth radiation budget data and climate research. Rev. Geophysics, 24, 439-468.

Hollingsworth, A., A.C. Lorenc, M.C. Tracton, K. Arpe, G. Cats, S. Uppala and P. Kållberg, 1985: The response of numerical prediction system to FGGE level IIb data. Part I: Analyses. Quart. J. Roy. Meteor. Soc., 111, 1-66.

Holopainen, E.O., 1968: A determination of the intensity of the atmospheric energy cycle from the vertical distribution of diabatic heating. Geophysica, 10, 69-73.

Holopainen, E.O. and K. Eerola, 1979: A diagnostic study of the long-term balance of kinetic energy of atmospheric large-scale motion over the British Isles. Quart. J. Roy. Meteor. Soc., 105, 849-858.

Holopainen, E.O., N-C. Lau and A.H. Oort, 1980: A diagnostic study of the time-averaged budget of atmospheric zonal momentum over North America. J. Atmos. Sci., 37, 2234-2242.

Holopainen, E.O. and C. Fortelius, 1986: Accuracy of the estimates of atmospheric large-scale energy flux divergence. Mon. Wea. Rev., 114, 1910-1921.

-"-, 1987: The damping of potential enstrophy in the large-scale transient eddies in the wintertime troposphere. J.Atmos. Sci., 44, 1973-1980.

Hoskins, B.J., 1985: The middle latitude transients in the FGGE year and their role in the seasonal mean momentum and heat budgets. GARP Special Report No. 42, II58-II72, WMO, Geneva.

Hoskins, B.J., M.E. McIntyre and A.W. Robertsen, 1985: On the use and significance of isentropic potential vorticity maps. Quart. J. Roy. Met. Soc., 111, 877-946.

Jakobowitz, H. and R.J. Tighe, 1984: Earth radiation budget derived from the Nimbus ERB Experiment. J. Geophys. Res., 89, D4, 4997-5010.

Johnson, D.R., R.D. Townsend and M-Y. Wei, 1985: The thermally coupled response of the planetary scale circulation to the global distribution of heat sources and sinks. Tellus, 37A, 106-125.

Kasahara, A. and A.P. Mizzi, 1985: Preliminary evaluation of diabatic heating distribution from FGGE level IIIb analysis data. Proc. of the First National Workshop on the Global Weather Experiment. National Academy Press, Washigton D.C., Vol.2, Part I, 317-329.

Kung, E.C. and W.E. Baker, 1975: Energy transformations in the middle latitude disturbances. Quart. J. Roy. Meteor. Soc., 101, 793-815.

Kung, E.C. and H. Tanaka, 1983: Energetics analysis of the global circulation during the Special Observing Periods of FGGE. J. Atmos. Sci., 40, 2575-2592.

Lau, N-C., 1979: The observed structure of tropospheric stationary waves and the local balance of vorticity and heat. J. Atmos. Sci., 36- 996-1016.

Lorenc, A.C. and R. Swinbank, 1984: On the accuracy of the general circulation statistics calculated from FGGE data - a comparison of results from two sets of analyses. Quart. J. Roy. Meteor. Soc., 110, 915-942.

Masuda, K., 1988: Meridional heat transport by the atmosphere and oceans: Analysis of FGGE data. To appear in Tellus.

Newell, R.E., J.W. Kidson, D.G. Vincent and G.J. Boer, 1972/1974: The General Circulation of the Tropical Atmosphere and Interactions with Extratropical Latitudes, Vols. I/II, The MIT Press, 258/371 pp.

Oort, A.H. and T.H. Vonder Haar, 1976: On the observed annual cycle in the ocean-atmosphere heat balance over the Northern Hemisphere. J. Phys. Oceanogr., 6, 781-800.

-"-, and J.P. Peixoto, 1983: Global angular momentum and energy balance requirements from observations. Advances in Geophysics, 25, 355-489.

Palmer, T.N., G.J. Shutts and R. Swinbank, 1986: Alleviation of systematic westerly bias in general circulation and numerical weather prediction models through an orographic gravity wave drag parametrization. Quart. J. Roy. Meteor. Soc., 112, 1001-1039.

Savijärvi, H., 1988: Global energy and moisture budgets from rawinsonde data. Mon. Wea. Rev., 115, xx-xx.

Swinbank, R., 1985: The global atmospheric angular momentum balance inferred from analyses made during FGGE. Quart. J. Roy. Meteor. Soc., 111, 977-992.

Townsend, R.D. and D.R. Johnson, 1985: A diagnostic study of the isentropic zonally averaged mass circulation during the First GARP Global Experiment. J.Atmos. Sci.,42, 1565-1579.

Uppala, S., 1986: Reassimilation of the final FGGE level IIIb dataset at ECMWF. Newsletter No. 14 of the U.S. Committee for GARP FGGE Advisory Panel, pp. 1-12.

Wiin-Nielsen, A. and J.A. Brown, 1962: On diagnostic computations of atmospheric heat sources and sinks and the generation of available potential energy. Proc. Int. Symp. on Numerical Weather Prediction. Tokyo, Meteorological Society of Japan, 593-613.