ON THE ROLE OF BAROTROPIC ENERGY CONVERSIONS IN THE GENERAL CIRCULATION

J. M. WALLACE
Department of Atmospheric Sciences
University of Washington
Seattle, Washington

AND

N.-C. LAU
Geophysical Fluid Dynamics Laboratory/NOAA
Princeton University
Princeton, New Jersey

1. Introduction ................................................. 33
2. Structure of the Transient Eddies. ....................... 36
3. A Vectorial Representation of the Barotropic Conversion 37
4. Observational Results .................................... 41
   4.1. NMC Analyses; Northern Hemisphere Winter ...... 41
   4.2. Station Data ......................................... 50
   4.3. Analyses from the Global Weather Experiment ...... 52
5. Results Based on GCM Simulations ....................... 55
6. Net Transient Eddy Forcing of the Mean Flow .......... 64
7. Further Interpretation .................................... 69
References .................................................. 72

1. INTRODUCTION

In the ongoing quest for an understanding of the atmospheric general circulation, two of the central problems are (1) to identify the source or sources of nonseasonal temporal variability and (2) to understand how this temporal variability feeds back on the seasonally varying, time-averaged circulation. These problems are recurrent themes in the modern general circulation literature dating back to the works of Rossby, Starr, and Charney in the 1940s.

As a result of the collective efforts of many investigators, it has become widely accepted that baroclinic instability is the main source of transient variability in the general circulation, and that in the later stages of their life cycle, baroclinic waves feed energy into the time-mean zonally averaged flow and planetary-scale low-frequency transients by barotropic exchange processes. This view is exemplified by Smagorinsky's (1972) juxtaposition of the
terms "barotropic stability" and "baroclinic instability" as section headings for one of his major review articles on general circulation research.

Among the more important pieces of evidence in support of this interpretation are

1. the rapid growth rates associated with baroclinic instability of realistic atmospheric flows (Charney, 1947; Eady, 1949; Frederiksen, 1982);

2. the fact that the structure and evolution of extratropical disturbances on daily weather maps strongly resemble those of baroclinic waves as simulated in nonlinear integrations based on a primitive equation model (Simmons and Hoskins, 1978; Hoskins, 1983a); these simulated waves give up much of their kinetic energy to the mean flow during the later stages of their life cycle [see Held and Hoskins (this volume)];

3. the dominance of the baroclinic conversion terms $C_A$ and $C_E$ in the Lorenz (1955) kinetic energy cycle as deduced from observations [e.g., Oort (1964) and Oort and Rasmusson (1971)] and general circulation model (GCM) simulations [e.g., see Phillips (1956) and Smagorinsky (1963)], together with the prevailing countergradient eddy transport of westerly momentum, which is evidence of a conversion of kinetic energy from the eddies into the zonally averaged flow;

4. the theoretical arguments of Onsager (1949), Fjortoft (1951), and others that two-dimensional turbulence should be characterized by a "de-cascade" of kinetic energy from the scale of the forcing (presumably baroclinic instability) toward larger scales; the predicted "de-cascade" has been verified in a number of observational studies [e.g., see the reviews of Saltzman (1970) and Holopainen (1983)].

The preceding evidence in favor of baroclinic instability/barotropic stability as a conceptual model for interpreting the energetics of large-scale atmospheric motion is based on theoretical analyses, modeling studies, and observational diagnostics that either explicitly or implicitly ignore the strong stationary wave patterns induced by the major mountain ranges and land–sea thermal contrasts. That the stationary waves might conceivably influence the structure and energetics of the transient eddies has long been recognized [e.g., see Saltzman (1963)], but it has only been within the past decade that this problem has attracted widespread interest.

The works of Blackmon (1976), Blackmon et al. (1977), and Lau (1978, 1979) established that the transient eddy variance and covariance statistics for fluctuations with periods on the order of a week or less are organized in terms of elongated "storm tracks" that lie slightly poleward of the climatological mean jet-exit regions over the oceanic sectors of the Northern Hemisphere during wintertime. Frederiksen (1979, 1982) was able to simulate many of the features in these distributions in a linear stability analysis of the
longitudinally dependent climatological mean flow. The effects of these transient eddy fluxes on the time-mean flow have been investigated by Hoskins (1983b) and by Lau and Holopainen (1984).

From the studies cited above it is evident that the stationary waves tend to organize baroclinic disturbances into preferred longitudinal sectors but do not fundamentally alter their wavelength, their meridional and vertical structure, their evolution, their fluxes in the meridional plane, or their energetics. Results of Hoskins (1983b) and Hoskins et al. (1983) indicate that in addition to their predominantly countergradient meridional fluxes of zonal momentum, longitudinally localized baroclinic waves produce a systematic westward flux of zonal momentum that is also countergradient, since the major storm tracks lie in the exit regions of westerly jetstreams. This westward flux enhances the barotropic energy conversion from the baroclinic waves into the time-mean flow as they evolve through their life cycle.

Fluctuations with periods longer than about 10 days exhibit a structure and evolution quite different from their higher-frequency counterparts. On these longer time scales the presence of the stationary waves is felt through their influence on the dispersion of trains of Rossby waves (Branstator, 1983) and through their role in organizing a significant fraction of the low-frequency variability into standing oscillations with geographically fixed nodes and antinodes (frequently referred to as "teleconnection patterns"). The first effect is most evident at intermediate time scales (10–30-day periods), while the second is most evident at long time scales (periods longer than a month and in the interannual variability of wintertime means) (Blackmon et al., 1984a,b).

The specific dynamical mechanism(s) responsible for organizing the low-frequency variability into teleconnection patterns are not yet well understood, but Simmons et al. (1983) have shown evidence that barotropic instability of the climatological mean flow plays a role in several of the observed patterns. In terms of structure and evolution, the normal modes associated with this "two-dimensional barotropic instability" are fundamentally different from the modes associated with the classic zonally symmetric barotropic instability problem (Kuo, 1949). The primary mechanism by which they extract kinetic energy from the zonal flow is through the eastward (down-gradient) flux of zonal momentum in the jet-exit regions. If this process proves to be important, then there will be a need for a reinterpretation of the role of barotropic conversion processes in the atmospheric general circulation.

In this chapter we will be concerned with the interactions between the transient eddies and the longitudinally dependent time-mean flow. After a brief review of the structure of the eddies, we will examine the barotropic conversion term in the Earth's atmosphere for both hemispheres and both
seasons. For the Northern Hemisphere winter we will show results based on three different observational data sets. We will contrast the observational results with comparable results based on three different GCM experiments carried out at the Geophysical Fluid Dynamics Laboratory (GFDL): one for an Earth-like planet; one for a planet with a land–sea distribution like the Earth’s, but with no mountains; and one for a planet with neither mountains nor oceans. Throughout the chapter we will be contrasting the behavior of high-frequency (periods less than 10 days) fluctuations and lower-frequency fluctuations, which appear to play very different roles in the general circulation.

2. Structure of the Transient Eddies

Following Hoskins et al. (1983), the velocity correlation tensor can be divided into isotropic and anisotropic (trace-free) components

\[
\begin{bmatrix}
K & 0 \\
0 & K
\end{bmatrix} + \begin{bmatrix}
M & N \\
N & -M
\end{bmatrix}
\]

(2.1)

where

\[K = (\overline{u^2} + \overline{v^2})/2, \quad M = (\overline{u^2} - \overline{v^2})/2, \quad N = u'v'\]

(2.2)

Here the overbars represent climatological (time) means and the primes represent deviations from them. The principal axis of the velocity correlation tensor is oriented at an angle

\[\theta = \frac{1}{2} \tan^{-1}(N/M)\]

(2.3)

relative to the \(x\) axis. In coordinates (\(\hat{x}, \hat{y}\)) aligned with the major and minor axes,

\[\hat{M} = (\overline{\hat{u}^2} - \overline{\hat{v}^2})/2, \quad \hat{N} = 0\]

(2.4)

Since \(K\) and \(\hat{M}\) are invariant under the rotation of the axes, it is possible to define a coefficient of anisotropy

\[\alpha = \hat{M}/K\]

(2.5)

which lies between zero and one, where \(\alpha \to 1\) as \(\overline{\hat{v}^2} \to 0\).

The line segments in Fig. 1 depict the minor axes of the velocity correlation tensor, calculated by applying the preceding relationships to National Meteorological Center (NMC) operational 300-mb height analyses for eight Northern Hemisphere winter seasons (November 15–March 14, 1966–1967 to 1968–1969 and 1970–1971 to 1974–1975). Data processing procedures are as described by Lau et al. (1981). We have chosen to display minor rather than major axes because observational results by Blackmon et
al. (1984b) indicate that over much of the hemisphere the group velocity associated with Rossby-wave dispersion tends to be eastward relative to the ground and parallel to the minor axes wherever it has a substantial zonal component. The length of the line segments is proportional to the coefficient of anisotropy.

The loops in Fig. 1 represent the 0.3 contour on selected one-point correlation maps for wintertime 500-mb height, as computed in Blackmon et al. (1984a). In most cases they resemble ellipses with minor axes corresponding closely to those associated with the velocity correlation tensor. The degree of ellipticity is related to the coefficient of anisotropy; where $\alpha$ is small, as indicated by short line segments, the loops tend to be more or less circular, whereas as $\alpha$ becomes larger, they become more like closed slits.

Figure 1a shows results based on unfiltered data, which exhibit only a modest amount of anisotropy, with $\alpha$ reaching values of 0.4 over the western United States. Even where $\alpha$ is small, there is a generally good agreement between the orientation of the closed curves and the minor axes of the velocity correlation tensor. Results for filtered data, shown in Fig. 1b,c, indicate a somewhat higher degree of anisotropy, with $\alpha$ reaching values of 0.55 over some regions in the low-pass filtered data.

Transient eddies in the 2.5- to 6-day period range [isolated by subjecting the wind and geopotential height data to the bandpass filter described by Blackmon (1976)] tend to be organized as east–west wave trains, in which individual disturbances have longer meridional scales than zonal scales. There is a prevailing tendency for a WNW–ESE orientation of the minor axes, particularly at the lower latitudes, which is indicative of an equatorward component of the group velocity. The bands of long line segments with similar alignment extending across the North Pacific and North America and across the North Atlantic and western Europe are suggestive of waveguides.

Eddies with periods longer than 10 days [isolated using the low-pass filter described by Blackmon (1976)] tend to be elongated in the east–west direction, as evidenced by the prevailing north–south orientation of their minor axes in Fig. 1c. Their anisotropy is particularly large over the oceanic sectors of the hemisphere. Over North America and parts of Eurasia there is evidence of the NW–SE Rossby-wave dispersion documented by Blackmon et al. (1984b).

3. A Vectorial Representation of the Barotropic Conversion

In Cartesian coordinates, the barotropic energy conversion from the transients into the time-mean flow is given by

$$ C = \overline{u'^2}(\partial \overline{u}/\partial x) + \overline{v'^2}(\partial \overline{v}/\partial y) + \overline{u'v'}(\partial \overline{v}/\partial x + \partial \overline{u}/\partial y) $$  \hspace{1cm} (3.1)
Fig. 1. Line segments indicate orientation of minor axes of local eddy correlation tensors; lengths are proportional to $\alpha = \hat{M}/K$, the coefficient of anisotropy. Scale given at lower left. Calculations based on eight winters of 300-mb NMC wind analyses, as described in text. Closed loops represent 0.3 contours of one-point correlation maps for 500-mb height at selected grid points, indicated by dots, based on data used in Blackmon et al. (1984a). (a) Unfiltered, twice-daily data; (b) bandpass (2.5–6-day period) filtered data; (c) low-pass (>10-day period) filtered data. Except for Figs. 5, 6c, and 6d, the outermost latitude circle for all polar stereographic plots in this chapter is 20°N. Latitude circles and meridians are drawn at intervals of 20°.

To an accuracy of about 10%, it can be assumed that the time-mean flow in extratropical latitudes is nondivergent, and $\partial v/\partial x$ is negligible in comparison to $\partial u/\partial y$, so that

$$C \equiv (u'^2 - \overline{v'^2})(\partial \overline{u}/\partial x) + u'\overline{v'}(\partial \overline{u}/\partial y)$$

(3.2)

or, following Hoskins et al. (1983) and Simmons et al. (1983),

$$C \equiv -\mathbf{E} \cdot \nabla \overline{u}$$

(3.3)

where

$$\mathbf{E} = -(\overline{u'^2 - v'^2}, u'v')$$

(3.4)
In defining \( \mathbf{E} \), we have followed the sign convention established by Hoskins et al. (1983), which is motivated by the Eliassen–Palm flux formulation of Edmon et al. (1980). With this definition, the \( \mathbf{E} \) vector tends to be in the direction of the relative group velocity, and the effective transient eddy flux of westerly momentum is directed in the opposite sense as \( \mathbf{E} \).

For \( M \gg N \), \( \mathbf{E} \) assumes a nearly zonal orientation. If \( \nu^2 > \bar{u}^2 \), \( \mathbf{E} \) is directed eastward, and since

\[
\tan^{-1}(N/2M) = \frac{1}{2} \tan^{-1}(N/M) \tag{3.5}
\]

it almost exactly coincides with the minor axis of the velocity correlation tensor, as indicated in Fig. 2a,b. Such a relationship is observed throughout most of temperate latitudes in the bandpass filtered data (Fig. 1b). For \( M \gg N \) and \( \nu^2 > \bar{v}^2 \), \( \mathbf{E} \) is directed westward, and for the same reason, it almost exactly coincides with the major axis of the velocity correlation tensor, as in Fig. 2e. Such a relationship is observed over the oceanic sectors in the low-pass filtered data (Fig. 1c). These situations were emphasized by Hoskins et al. (1983). It should be noted that if \( N \) is larger than \( M/2 \), \( \mathbf{E} \) will not be parallel to the axes of the eddy correlation tensor; e.g., see Fig. 2c,d,f.

In contrast to the axes of the velocity correlation tensor, which are invariant under rotation of the coordinate system, \( \mathbf{E} \) is a pseudovector whose orientation depends on the coordinate system in which it is computed; e.g., the \( \mathbf{E} \) field computed in natural coordinates oriented relative to the local absolute vorticity contours is noticeably different from that computed in conventional latitude–longitude coordinates [see Hoskins et al. (1983)].
4. Observational Results

4.1. NMC Analyses; Northern Hemisphere Winter

Figure 3a shows the distributions of $E$ and $\vec{u}$ based on the same data set as Fig. 1. It is evident that the distribution of $E$ and its relation to the zonal wind field is rather complicated and exhibits a strong regional dependence. Large conversions of kinetic energy from the time-mean flow to the transient eddies, as evidenced by $E$ vectors directed up the gradient of $\vec{u}$, occur in the jet-exit region over the central Pacific and to the north of the jet-entrance regions over the southern United States and North Africa. Conversions in the reverse sense occur over much of the Atlantic. These areas correspond to extrema in the distribution of $E \cdot \nabla \vec{u}$, shown in Fig. 4a.

In the bandpass (2.5- to 6-day period) filtered data, shown in Fig. 3b, the most striking features are the bundles of eastward $E$ vectors, directed down the gradient of $\vec{u}$ in the major storm tracks located slightly poleward of the jet-exit regions over the oceanic sectors of the hemisphere. Blackmon et al. (1977, 1984a) and Lau (1979) have argued that the structure of the high-frequency transient eddies within these storm-track regions can be identified with baroclinic wave activity. As in the unfiltered pattern, $E$ is directed up the gradient of $\vec{u}$ over much of the United States. The pattern of $E \cdot \nabla \vec{u}$, shown in Fig. 4b, is dominated by regions of strong conversion of kinetic energy from the transient eddies into the mean flow over the oceanic sectors that are not so apparent in the unfiltered data (Fig. 4a). In these regions, the eddies with short time scales transport westerly momentum westward, toward the jet streams, thereby increasing the kinetic energy of the time-mean flow at the expense of the transient eddies.

Figure 3c shows a sharply contrasting picture for the low-frequency component of the transient eddies, which corresponds to fluctuations with periods longer than 10 days. Throughout most of the hemisphere, the $E$ vectors have a westward component consistent with the east–west elongation of the low-frequency eddies in Fig. 1c. The longest $E$ vectors in Fig. 3c are located in the jet-exit region over the central Pacific, where they are directed up the gradient of $\vec{u}$. An analogous but weaker region of westward $E$ vectors is located over the Atlantic jet-exit region. The same features show up with greater clarity in monthly mean data (Fig. 3d), which emphasize transient fluctuations with periods on the order of 60 days or longer and the interannual variability of wintertime means. The westward $E$ vectors in the jet-exit regions are indicative of a barotropic energy conversion from the time-mean flow into the low-frequency transient eddies. These features are evident in the corresponding distributions of $E \cdot \nabla \vec{u}$, shown in Fig. 4c,d,
Fig. 3. E vectors, as defined in Eq. (3.4), superimposed on isotachs of the zonal component of the climatological mean flow based on eight winters of NMC 300-mb data for Northern Hemisphere winter. Contour interval for zonal wind 5 m s$^{-1}$; scales for E vectors given at lower left. (a) Unfiltered data, (b) bandpass (2.5–6-day period) filtered data, (c) low-pass filtered (>10-day period) data, and (d) monthly mean data. Scale for (b)–(d) is one-half that of (a). Panels (a), (b), and (c) are based on data for November 15–March 14; (d) is based on data for November, December, January, February, and March (i.e., 5-month winters, 40 months altogether).
FIG. 4. The barotropic energy conversion from the climatological mean flow into the transient eddies, as inferred by evaluating $E \cdot \nabla \bar{u}$ from the 300-mb data in Fig. 3. The shaded regions denote positive conversions (i.e., conversions in the opposite sense as in the Lorenz kinetic energy cycle). The zero contour interval is omitted for the sake of clarity. Contour interval for (a) is $4 \times 10^{-4} \text{ m}^2 \text{s}^{-3}$ and for (b)-(d) is $2 \times 10^{-4} \text{ m}^2 \text{s}^{-3}$. (a) Unfiltered data, (b) bandpass data, (c) low-pass data, (d) monthly means.
and they are responsible for some of the main features in the unfiltered data (Fig. 4a).

It is interesting to note the similarity between the distribution of $E$ in Fig. 3d and the theoretically calculated distribution for the fastest-growing normal mode associated with barotropic instability of the climatological mean January 300-mb flow from Simmons et al. (1983), which is reproduced in Fig. 5. The strong correspondence supports their contention that two-dimensional barotropic instability is an energy source for low-frequency transient fluctuations in Northern Hemisphere winter. As further evidence,
Simmons et al. (1983) note that

(1) the fastest-growing normal mode associated with barotropic instability resembles two observed teleconnection patterns, as illustrated in Fig. 6;

(2) the "centers of action" in the stream function of this same fastest-growing normal mode correspond closely to the regions of strongest low-frequency variability of the geopotential height field, which lie over the northern oceans [e.g., see Blackmon et al. (1984a)];

(3) the normal modes associated with barotropic instability of the time-mean flow are characterized by much-lower-frequency oscillatory components than those associated with baroclinic instability; e.g., the fastest-growing normal mode has a period on the order of 50 days. This frequency separation between the barotropic and baroclinic normal modes is consistent with the distinctions between Fig. 3b and Fig. 3c,d.

Table I shows the hemispheric average (poleward of 20°N) of the conversion of kinetic energy from the transient eddies into the climatological mean flow, as defined in Eq. (3.1) and including the terms associated with the earth’s spherical geometry, at the 300-mb level. There is a strong compensation between the high- and low-frequency fluctuations, as evidenced by the fact that the conversions for the time-filtered eddies are about an order of

| Table I. Barotropic Energy Conversions at the 300-mb Level for Northern Hemisphere Winter^a |
|----------------------------------|--------------|--------------|--------------|
|                                  | NMC Data     |              |              |
|                                  | Bandpass (2.5–6-day period) | Low pass (>10-day period) | Unfiltered |
| C                                | 3.8          | -3.7         | -0.3         |
| \( K_T \)                        | 16           | 43           | 108          |
| \( C_K \)                        | 0.5          | 1.4          | 3.5          |

<table>
<thead>
<tr>
<th></th>
<th>Mountain GCM Experiment</th>
</tr>
</thead>
<tbody>
<tr>
<td>C</td>
<td>2.2</td>
</tr>
<tr>
<td>( K_T )</td>
<td>13</td>
</tr>
<tr>
<td>( C_K )</td>
<td>0.6</td>
</tr>
</tbody>
</table>

^a C is as defined in Eq. (3.1) including the terms associated with the earth's spherical geometry, and \( K_T \) represents the kinetic energy of the transient eddies. \( C_K \) is the conversion of kinetic energy from the eddies into the zonally averaged flow, as defined in the Lorenz (1955) kinetic energy cycle. Kinetic energy (per unit mass) is given in units of square meters per square second; conversions in units of \( 10^{-5} \text{m}^2\text{s}^{-3} \) (approximately equivalent to square meters per square second per day).
Fig. 6. (a) and (b) Sample teleconnection patterns based on NMC monthly mean 500-mb height data for Northern Hemisphere winter. [From Wallace and Gutzler (1981). Reproduced with permission from *Monthly Weather Review*, a publication of the American Meteorological Society.] (c) and (d) Sample geopotential height patterns obtained by Simmons *et al.* (1983) in numerical integrations based on the barotropic vorticity equation, perturbed about the 300-mb January climatological mean flow. [From Simmons, Wallace, and Branstator (1983). Reproduced with permission from *Journal of the Atmospheric Sciences*, a publication of the American Meteorological Society.] In both cases the flow was perturbed by inserting a local disturbance of very small amplitude into the initial conditions; the maps shown here represent the simulated 300-mb height perturbations 6 days later. In (c) the initial disturbance was centered on 30°N, 120°E, and in (d) it was centered on 15°N, 90°W. Patterns similar to (a) and (b) can be obtained from day 6 of the integrations onward in experiments with the initial disturbance placed in a wide variety of different locations.
magnitude larger than the net rate of conversion for the transient eddies as a whole. Comparing the top two rows of Table I, it is evident that the high-frequency transients feed energy into the mean flow at the jet-stream level at a rate sufficient to exhaust themselves on a time scale of 4 to 5 days on a hemispheric basis. However, it should be borne in mind that the conversion rates in Table I represent averages over regions of contrasting sign. Locally, in the storm tracks, the barotropic decay time is on the order of 1 to 2 days, and over the central United States the barotropic amplification time is equally short. On a hemispheric basis, roughly half the high-frequency barotropic conversion is due to the meridional term $u'v' \partial \bar{u}/\partial y$ in Eq. (3.2) and the other half is due to the term $(\bar{u}_x^2 - \bar{v}_x^2) \partial \bar{u}/\partial x$.

The low-frequency transients extract energy from the mean flow at a rate sufficient to replenish themselves on a time scale of about 2 weeks. It is evident from Figs. 3c and 4c that most of this conversion takes place in the jet-exit regions, but the poleward flanks of the jet-entrance regions over the central United States and the Mediterranean also make significant contributions to the hemispheric average. The term $(\bar{u}_x^2 - \bar{v}_x^2) \partial \bar{u}/\partial x$ dominates the low-frequency barotropic conversion.

4.2. Station Data

Over data sparse regions where the NMC wind fields are influenced by the procedures employed in interpolating irregularly spaced station data to grid-point fields, it is reasonable to question the validity of the features in the $E$ and $\bar{u}$ fields described in Figs. 3 and 4. Following the approach in an investigation by Lau and Oort (1981), we present in Fig. 7a distributions of $E$ and $\bar{u}$ based on a 10-winter (1963–1964 through 1972–1973) station data set compiled by Oort (1983) at the GFDL. The variances and covariances in Oort’s data set are based on deviations from means for individual year/months, and therefore the contribution from the very-low-frequency variability is missing from these statistics. Nevertheless, they should be roughly comparable to the unfiltered results in Fig. 3a. Figure 7b shows $E$ and $\bar{u}$ fields derived from monthly mean wind statistics for selected stations tabulated in Monthly Climatic Data for the World (World Meteorological Organization, 1963–1973), the counterpart of Fig. 3d. To the extent that they can be compared, Figs. 3 and 7 appear to be at least qualitatively consistent with respect to the major features in the $E$ field. In particular, Fig. 7 provides a confirmation of the westward-directed $E$ vectors associated with low-frequency transient disturbances in the jet-exit regions in the oceanic sectors and the southeastward-directed arrows on the poleward flanks of the jet-entrance regions over the United States and the Mediterranean region. Hence there can be little doubt concerning the reality of these features.
Fig. 7. \( \mathbf{E} \) vectors and \( \bar{u} \) contours based on station data; contour intervals 5 m s\(^{-1}\). (a) Oort's 10-year data set (1963–1973) in which \( \mathbf{E} \) vectors are associated with fluctuations of daily values about means for individual year/months; these results should be roughly comparable to the unfiltered results in Fig. 3a. (b) Selected station data extracted from Monthly Climatic Data for the World (World Meteorological Organization) also for the period 1963–1973, comparable to the monthly mean results in Fig. 3d. The scale for the arrows in (b) is expanded by a factor of two relative to that of Fig. 3d. The contours of \( \bar{u} \) are based on Oort's 10-winter (1963–1973) analysis.
4.3. Analyses from the Global Weather Experiment

A major task carried out at the GFDL in connection with the Global Weather Experiment was the production of the "level III-B data set" consisting of analyzed fields based on the full set of satellite and ground-based observations taken during the experiment and transmitted to the World Data Centers, either in real time or in the months that followed the experiment [see Ploshay et al. (1983) for details].

Fields of $E$ and $\bar{u}$ derived from this data set for December–February 1978–1979 and June–August 1979 are shown in Figs. 8 and 9. The field for Northern Hemisphere winter (Fig. 8a) exhibits all the major features described in the preceding section. Hence a single year's data should be sufficient for describing the relation between $E$ and $\bar{u}$ on a global basis. In comparison to their wintertime counterparts, the summertime $E$ vectors and the gradients of $\bar{u}$, shown in Fig. 8b, are weaker and there is less large-scale organization in the $E$ field. The eastward-directed arrows across northern Europe and Asia are suggestive of a storm track. There are westward directed arrows in the shear zones to the south of the mid-latitude jet streams.

The distributions for the Southern Hemisphere, shown in Fig. 9, both exhibit a considerable amount of structure. Prominent features in the wintertime distribution (Fig. 9b) are the eastward-directed bundle of $E$ vectors emerging from the exit of the high-latitude jet stream over the Indian Ocean and the equatorward vectors on the poleward flank of the jet-entrance region over and to the west of Australia. Comparison with results of Hoskins et al. (1983) (their Fig. 11), based on the 1980 winter, tends to confirm the existence of these features. The former appears to be associated with the primary storm track in the hemisphere and the latter is perhaps an analog to the wintertime jet-entrance regions over the southern United States and North Africa.

The summertime distribution (Fig. 9a) is dominated by a ring of eastward-directed $E$ vectors that coincides with the strongest westerlies, near 50°S. At a number of different longitudes, bundles of arrows peel off from the equatorward flank of the jet and curve northward into the subtropics.

It is interesting to note that of the four distributions shown in Figs. 8 and 9, only the Northern Hemisphere winter (Fig. 8a) shows evidence of westward $E$ vectors in the jet-exit regions. There are several examples of bundles of eastward-pointing vectors that curve equatorward on the poleward flanks of jet-entrance regions. Apparently, such regions account for most of the observed poleward flux of westerly momentum in the 30 to 40° latitude belt.
Fig. 8. As in Fig. 3a, but for the Global Weather Experiment [First GARP (Global Atmospheric Research Programme) Global Experiment (FGGE)] Level III-B data set (E vectors, $\bar{u}$ contours; 300-mb unfiltered data). (a) December 1978 – February 1979, (b) June – August 1979. Shading indicates easterlies.
5. Results Based on GCM Simulations

In 1978–1980 the Climate Dynamics group at the GFDL carried out a pair of 15-year integrations with a low-resolution spectral model. [For details of the model formulation, see Gordon and Stern (1982).] Following the experimental strategy used by Manabe and Terpstra (1974), one of the integrations was carried out with realistic boundary conditions at the Earth’s surface, and the other was identical but for the fact that the orography was removed. The two integrations are referred to as the “mountain (M)” and the “no mountain (NM)” experiments, respectively. In this section we shall describe results for the Northern Hemisphere during winter only. The results described in this section are averages based on the first 10 years of 15-year integrations.

From an inspection of the hemispherically averaged energetics statistics shown in Table I, one might easily conclude that the M experiment did not produce a very good simulation of the observed kinetic energy cycle. The transient eddy kinetic energy is too small by more than a factor of two, and the conversion between the transient eddies and the climatological mean flow is an order of magnitude larger than the observed and in the opposite direction. Even the term \( (u'^2 - v'^2)(\partial \overline{u}/\partial x) \) (not shown) is from the eddies into the mean flow, whereas in the NMC data it is strong and in the reverse sense. However, a comparison of the distributions of \( E \) and \( \overline{u} \), shown in Fig. 10, and \( E \cdot \nabla \overline{u} \), shown in Fig. 11, with the corresponding distributions based on NMC data shown in Figs. 3 and 4, respectively, gives rather convincing evidence that the GCM is capable of simulating the barotropic interactions between the mean flow and the eddies at the jet-stream level at least qualitatively, if not in quantitative detail. This conclusion is reinforced by a comparison of Figs. 12 and 1, which shows that the horizontal structures of the transient eddies in the GCM simulation are realistic in terms of scale and shape. The marked differences in the hemispherically averaged barotropic energy conversions between the M experiment and the NMC data in Table I are evidently a reflection of the differing relative strengths of the various regional features.

The “no mountain” experiment yielded a jet-stream structure more reminiscent of the Southern Hemisphere. The \( E \) vectors shown in Fig. 13 are predominantly eastward in the latitude belt of the jet stream. Equatorward of the jet stream they peel off equatorward and weaken. Within the primary jet-entrance region over Eurasia, the \( E \) vectors stream southeastward on the poleward flank of the jet, where they cross the isolochs toward higher values of \( \overline{u} \). A similar but weaker region of southeastward pointing \( E \) vectors occurs over eastern Canada, poleward of the weak secondary jet maximum off
Fig. 10. As in Fig. 3, but for the first 10 winters of the mountain GCM experiment. Scale for arrows is expanded by 50% relative to that of Fig. 3. (a) Unfiltered data, (b) bandpass data, (c) low-pass data, (d) monthly means.
Fig. 11. As in Fig. 4, but for the first 10 winters of the mountain GCM experiment. (a)–(d) as in Fig. 10. The contour interval is one-half that of Fig. 4, i.e., (a) $2 \times 10^{-4} \text{ m}^2 \text{s}^{-3}$, (b)–(d) $1 \times 10^{-4} \text{ m}^2 \text{s}^{-3}$. 
Nova Scotia. These regions are characterized by strong poleward, down-gradient fluxes of westerly momentum.

The very-low-frequency variability, as represented by monthly mean data shown in Fig. 13b, contributes very little to the poleward flux of westerly momentum. The main form of anisotropy at these very low frequencies is the prevailing westward sense of the arrows, which indicates that $u'^2 > v'^2$ and that the disturbances are elongated in the east–west direction. The strongest anisotropy is observed, not in the jet-exit regions, as in Fig. 3c,d, but in the shear zones to the north and south of the jet throughout most of the western half of the hemisphere. In the mountain experiment (Fig. 10c,d), there is a tendency for the westward E vectors in the jet-exit regions to bifurcate and flow around the jets so that the strongest anisotropy also lies in the shear zones, to the north and to the south. This tendency is particularly pronounced in the Atlantic sector.

The anisotropy in the shear zones is related to the distribution of telecon-
Fig. 13. As in Fig. 3, but for the no mountain GCM experiment. (a) Unfiltered data, (b) monthly mean data. Scale for arrows is expanded by 50% relative to that of Fig. 3.
nectivity* shown in Fig. 14c, which is characterized by an elongated maximum extended along the axis of the jet stream, in sharp contrast to the dipole patterns straddling the jet-exit regions suggested by the NMC data shown in Fig. 14a. In teleconnection patterns with centers of action centered on the axis of the jet stream, the fluctuations in zonal wind have odd symmetry about the axis of the jet stream; i.e., positive deviations of $\bar{u}$ to the north of the jet axis are accompanied by negative deviations to the south of the jet axis and vice versa. Note that the pattern of teleconnectivity in the M experiment (Fig. 14b) also shows a tendency for maxima centered on the axis of the jet stream, rather than along the flanks of it as in the NMC data over the Atlantic (Fig. 14a). These differences are also reflected in the distributions of $\mathbf{E} \cdot \nabla \bar{u}$ (compare Figs. 4d and 11d).

A third GCM experiment was conducted with the same spectral model in which the lower boundary was prescribed to be a flat, moist land surface, and the solar flux was prescribed to be constant and equal to its annual mean value rather than to be seasonally varying as in the M and NM experiments. The history tapes from a 48-month integration with this “flat land” (FL) model integration have not yet been fully analyzed, but I. M. Held of the GFDL has kindly provided us with monthly mean data. The mean zonal wind distribution, indicated by the contours in Fig. 15, is nearly zonally symmetric. The small departures from zonal symmetry are a result of the finite sampling interval. Because of the annual mean radiative forcing, the jet stream is weaker and more diffuse than in the NM experiment, which was based on wintertime forcing. Maximum wind speeds are only on the order of 25 m s$^{-1}$. The distribution of $\mathbf{E}$ based on monthly mean data, indicated by the arrows in Fig. 15, differs from that in the NM experiment (Fig. 13b): whereas the anisotropy vanishes in the core of the jet stream in the NM experiment, $\mathbf{E}$ reaches its largest values in the jet stream in the FL experiment. Throughout the hemisphere $u'^2 > v'^2$, and the low-frequency eddies produce a weak poleward flux of westerly momentum.

In addition to the distributions of $\mathbf{E}$ and $\bar{u}$ displayed in Figs. 10 and 13, we have examined distributions for the Southern Hemisphere summer from both the M and NM experiments, which provide further examples of the kinds of structures described in this section. Results for the M experiment (not shown) resemble the pattern in the data from the Global Weather Experiment (Fig. 9a).

*The teleconnectivity at a given grid point is defined as the largest negative correlation coefficient that appears on the one-point correlation map for that grid point. It may also be defined as the column matrix whose elements are comprised of the largest negative correlation coefficients in each row (or column) of the Hermitian matrix whose elements are the correlation coefficients between the variable in question (in this case 500-mb height) at each grid point and each other grid point. For further details and examples, see Wallace and Gutzler (1981).
Fig. 14. The distribution of teleconnectivity of the 500-mb height field [as defined by Wallace and Gutzler (1981)] for the Northern Hemisphere winter, based on monthly mean data. (a) NMC data, (b) the mountain GCM experiment, and (c) the no mountain GCM experiment. Contour interval 0.1. Only values greater than 0.6 are contoured. Shading indicates values greater than 0.7.

6. Net Transient Eddy Forcing of the Mean Flow

It is of interest to determine whether the signature of any of the dynamical processes described in the previous sections is recognizable in the net forcing of the time-mean flow by the transient eddies. Lau and Holopainen (1984) have shown that the quasi-geostrophic tendency equation for the time-mean flow provides a convenient framework for inferring the net effect of the horizontal eddy fluxes of momentum and heat, taking into account the ageostrophic circulations that they induce. In this formulation, the convergence of the eddy fluxes plays a role analogous to the advection terms in the instantaneous tendency equation.

Figure 16, from Lau and Holopainen (1984), shows the tendencies in 300-mb geopotential height induced by the vorticity fluxes associated with high- and low-frequency fluctuations as defined earlier. The contour interval between solid contours is roughly equivalent to 5 m (of 300-mb height) per
Fig. 15. As in Fig. 3d, but for the flat land GCM experiment. Pressure level 202 mb. Scale for arrows is expanded by 50% relative to that of Fig. 3d.

day. Calculations are based on the NMC data set for Northern Hemisphere winter. The vorticity fluxes by the high-frequency transients (Fig. 16a) tend to accelerate the climatological mean jet streams along their northern flanks. Throughout the Pacific and American sectors of the hemisphere, the eddies produce westerly accelerations in the 40 to 50°N latitude belt. The low-frequency eddies (Fig. 16b) produce stronger accelerations of the mean flow, particularly over the eastern oceans. They tend to weaken the longitudinal contrasts in jet stream structure, accelerating the zonal flow where it is weak over the eastern Pacific and Europe, decelerating the Asian jet stream southeast of Japan, and shifting the jet streams over the western United States and North Africa poleward, bringing them more closely into line with the other jets. The total 300-mb height tendency induced by the transient eddies, shown in Fig. 17, closely resembles that associated with the vorticity flux induced by the low-frequency eddies (Fig. 16b). Hence, low-frequency barotropic processes dominate the wave–mean-flow interaction at the jet stream level.
Fig. 16. Distributions of the 300-mb geopotential tendencies associated with transient eddy transport of vorticity, for (a) bandpass (2.5- to 6-day period) and (b) low-pass (>10-day period) NMC data. Interval between solid contours $5 \times 10^{-4}$ m$^2$ s$^{-3}$. Arrowheads indicate the direction of geostrophic wind tendencies. [From Lau and Holopainen (1984). Reproduced with permission from Journal of the Atmospheric Sciences, a publication of the American Meteorological Society.]
Hoskins et al. (1983) have shown that to within about 10%, the barotropic component of the eddy forcing of the mean zonal flow is given by $\nabla \cdot E$. From a comparison of Figs. 3 and 16 it can be seen that all the features described in the preceding paragraph correspond to well-defined regions of divergence or convergence in the $E$ field. Hence the patterns in the $E$ vectors are helpful in interpreting the net eddy forcing of the mean zonal flow at the jet stream level.
7. Further Interpretation

All the distributions that we examined showed a tendency for eastward-directed $E$ vectors at the high frequencies and westward-directed $E$ vectors at the low frequencies, consistent with results of Hoskins et al. (1983) and Blackmon et al. (1984a,b). Hence, the filtered data exhibit considerably more anisotropy than do the unfiltered data. Much of this "polarization" of transient eddy structure in the frequency domain is due to differential doppler shifting associated with the time-mean flow and differential westward Rossby-wave propagation associated with the $\beta$ effect.

To describe the doppler-shifting phenomenon, let us consider the hypothetical situation of a horizontal wind field corresponding to homogeneous, isotropic turbulence superimposed on a mean zonal flow that, for the sake of argument, can be presumed to be a pure superrotation. For the present, let us ignore the influence of the $\beta$ effect in retarding or reversing the eastward advection of the larger-scale eddies. In such a flow, eddies of any given two-dimensional scale that are elongated in the east–west direction will require a longer time to be advected past fixed reference points than do eddies of the same scale that are elongated in the north–south direction. Alternatively, it can be argued that spherical harmonics of any given two-dimensional wave number $n$ that are elongated in the east–west direction will exhibit lower frequencies at fixed reference points than those elongated in the opposite sense, even though all the harmonics are subject to the same eastward advective phase speed.

The argument can be extended to include the influence of the $\beta$ effect as follows. For any given superrotation, there is a corresponding eddy scale or two-dimensional wave number $n$ for which the westward phase propagation induced by the $\beta$ effect just balances the eastward advection by the mean flow so that the wave is stationary relative to the ground. Surrounding this stationary wave number there exists a finite band of two-dimensional wave numbers for which the zonal phase speeds relative to the ground are smaller than any prescribed lower limit. It is these slow-moving waves that account for the patterns in the low-pass filtered and monthly mean data. However, it is not the phase speed alone that determines the ground-based frequency of the waves: it is the product of the phase speed and the zonal wave number. Hence, waves of a given two-dimensional wave number $n$ that are elongated in the zonal direction have lower ground-based frequencies than those that are elongated in the meridional direction. It follows that low-pass filtered data should contain a disproportionately large share of waves or eddies that are zonally elongated, and vice versa.

Apart from the FL experiment, all the other distributions of $E$ described in the preceding section exhibit spatially dependent anisotropy with distinctive horizontal structures that are evident even in the unfiltered data. These
regional features give rise to local convergence and divergence in the $E$ field, which is indicative of eddy forcing of the time-mean flow. They are associated with at least four distinct types of organization of the transient eddies:

(1) Bands of baroclinic wave activity, evident in the high-frequency component of the transient eddies: In the Northern Hemisphere winter, where the climatological mean jet streams have well-defined entrance and exit regions, baroclinic activity tends to be concentrated on the poleward flank of the jet-exit regions, where the $E$ vectors exhibit a diffluent pattern as they stream eastward, down the gradient of $u$, indicating that the eddies are giving up their kinetic energy to the mean flow (e.g., see Fig. 3b).

(2) The signature of two-dimensional barotropic instability characterized by westward, up-gradient $E$ vectors in the jet-exit regions at the low frequencies: This phenomenon occurs regularly in the Northern Hemisphere winter (Figs. 3d and 8a) and less strongly in the M experiment (Fig. 10d), but not in the Southern Hemisphere nor in the Northern Hemisphere summer nor in the NM or FL experiments; apparently it is important only in the presence of rather strong zonal gradients of the time-mean flow. When it occurs, it is capable of producing barotropic energy conversions comparable to (1), but in the reverse sense.

(3) Locally strong equatorward Rossby-wave dispersion on the poleward flanks of the jet-entrance regions as evidenced by the curving of the $E$ vectors and the orientations of the wave trains for the high- and intermediate-frequency transient eddies: The equatorward dispersion is indicative of a locally strong poleward momentum flux. Lau et al. (1978) found local maxima in the distribution of $u'v'$ in the same regions. These fluxes are down the gradient of westerly momentum, and thus they are indicative of a conversion of kinetic energy from the climatological mean flow into the transient eddies. The magnitude of the conversion appears to be comparable to that associated with (2), and this phenomenon appears to be more widespread than (2) (e.g., see Figs. 3a, 9a, 10a, and 13a). The only example of a jet-entrance region where there is not a strong conversion in this sense is over China and Mongolia during wintertime, and that is at least partially due to the low eddy kinetic energy levels in that region. Comparison of panels (b) and (c) in Figs. 1, 3, and 10 suggests that the equatorward refraction of the Rossby-wave trains is more pronounced for the low-frequency than for the high-frequency eddies, but it is evident from Figs. 4 and 11 that eddies with a broad range of frequencies contribute to the barotropic energy conversion in these regions; hence, the signature of this process is clearly evident in the unfiltered data. Poleward of $30^\circ$ latitude, most of the poleward flux of zonal momentum appears to be associated with these longitudinally dependent signatures. In contrast, there appears to be more widespread equatorward Rossby-wave
dispersion equatorward of $30^\circ$N on the anticyclonic side of the major jet streams.

(4) Bands of low-frequency zonal wind fluctuations centered in the shear zones to the north and south of elongated jet streams: This form of organization, in and of itself, produces little barotropic energy conversion, since the $\mathbf{E}$ vectors tend to be parallel to the $\bar{u}$ contours. It is most evident in the NM experiment, but there are hints of it in some of the other distributions, e.g., in the M experiment (Fig. 10), in the fastest-growing mode of barotropic instability (Fig. 5), and in the data from the Global Weather Experiment for the Northern Hemisphere summer (Fig. 8b). Teleconnection patterns in the geopotential height field associated with (2) are characterized by dipole structures straddling the jet-exit regions, whereas those associated with (4) are centered on the jet stream, with $u'$ showing a negative correlation between the poleward and equatorward flanks of the jet stream. From a synoptic point of view, (2) can be interpreted as an alternating lengthening and retraction of the jet streams in the downwind direction, whereas (4) can be interpreted as a north–south translation of the jets.

The preceding results in no way negate the importance of the conversion of kinetic energy from baroclinic waves into the climatological mean flow. The predominance of the pattern (1) in the high-frequency transient mean flow is evidence that this process plays an important role in the general circulation. However, it is clear that the barotropic energy conversion involves more than a single dynamical process. The patterns (2) and (3) represent other kinds of interactions between the eddies and longitudinally dependent mean flows that involve energy conversions from the mean flow into the transient eddies; two-dimensional barotropic instability (2) involves the very low frequencies and equatorward wave dispersion (3) involves a wide range of frequencies.

It is evident that the process (1) contributes to the maintenance of the climatological mean jet streams and to the general poleward flux of zonal momentum from the subtropics into the middle latitudes. However, it should be emphasized that these contributions are small. In contrast, the net effect of the processes (2) and (3), which dominate the eddy forcing of the mean flow at the jet-stream level, is to reduce the longitudinal contrasts in the time-mean flow pattern: (2) by reducing the zonal gradients in zonal wind speed across the jet-exit regions and (3) by extracting westerly momentum from the jet-entrance regions, near $30^\circ$N, and depositing it farther poleward in regions of relatively weak zonal winds. Hence, in accord with the results presented in Fig. 17, the eddies are tending to create a more zonally symmetric flow pattern with a westerly jet stream poleward of its observed position.
The demonstrated importance of these low-frequency processes in the interactions between the transient eddies and the time mean flow supports the case for broadening our conceptual understanding of the general circulation to encompass not only the time-mean flow and baroclinic waves, but also a third entity or group of entities, labeled "low-frequency fluctuations," that might well exist even in the absence of baroclinic instability. In contrast to baroclinic waves, which give up much of their kinetic energy to the barotropic component of the time-mean flow, these low-frequency fluctuations apparently feed on the time-mean flow and thereby serve to damp the longitudinally dependent, climatological-mean features at the jet-stream level.

Acknowledgments

We would like to thank Isaac M. Held, Eero O. Holopainen, and Abraham H. Oort for reviewing the paper and offering helpful suggestions; the GFDL Scientific Illustration Group for drafting the figures; and John Conner for supplying the photography. This work was supported by the Climate Dynamics Program of the National Science Foundation under Grant ATM 78-06099 and by the Geophysical Fluid Dynamics Program under NOAA Grant 04-7-022-44017.

References


