A Diagnostic Study of the Time-Averaged Budget of Atmospheric Zonal Momentum over North America

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ABSTRACT

The terms in the time-mean zonal momentum equation which depend only on the large-scale motions are evaluated for the North American continent during the winter season. The computations are based on two different sets of upper air atmospheric circulation statistics between 850 and 200 mb. The first (GFDL) data set consists of global objective horizontal analyses of monthly circulation statistics evaluated at individual stations; the second (NMC) data set was compiled by processing twice-daily synoptic analyses on a hemispheric grid. The mean residual forces needed for balance are discussed assuming that they represent the effects of horizontal and vertical subgrid-scale processes.

The results derived from the two largely independent data sets are similar. Zonal momentum is produced in the free atmosphere over North America by a local, thermally direct meridional circulation, with mean poleward ageostrophic flow above 700 mb. This production is partially counterbalanced by a net export of zonal momentum from the region. The flux divergence associated with large-scale vertical transports appears to be insignificant.

A residual force of considerable magnitude is needed for balance. In the lower troposphere, this force has an accelerating effect on the zonal flow almost everywhere over the eastern portion of the North American continent, and a decelerating effect over the western mountainous region where, however, the results are less certain. In the upper troposphere, the residual force over the entire continent has a net decelerating effect, but significant geographical differences appear to occur.

A tentative interpretation of the results for the residual force is offered.

1. Introduction

The first-order picture of how the balance of zonal momentum (or zonal angular momentum) is fulfilled in the earth's atmosphere has already been known for a few decades. The relative role played by the mean meridional circulation and atmospheric large-scale eddies in the maintenance of the zonally averaged circulation was of primary concern in various observational studies. The most recent comprehensive works in this area include those by Oort and Rasmusson (1971) and Newell *et al.* (1972, 1974). Recent investigations by Blackmon *et al.* (1977) and Lau (1978), among others, have shown how the first-order components of the momentum budget vary with longitude.

One of the still uncertain aspects of the atmospheric momentum balance is the forcing effect of subgrid-scale phenomena on the large-scale flow. Several studies seem to support the idea that "cumulus friction" is important in the momentum balance in the tropics (Schneider and Lindzen, 1976; Stevens *et al.*, 1977), and other observational evidence (Houze, 1973; Sanders and Emanuel, 1977; Vincent and Schlatter, 1979) suggests that cumulus convection may also be important in middle latitudes. However, these findings are not universally found to be convincing (Thompson and Hartmann, 1979).

One empirical method for studying the effects of subgrid-scale processes is to evaluate those terms in the budget equation which depend on the observed large-scale motions, and then to determine these effects as the residual. This approach was first used by Kurihara (1960) for Japan, and applied more recently by Holopainen (1979, hereafter referred to as H) for the British Isles. In the present paper a similar analysis is made during the winter season for North America, which is one of the largest regions with a good network of aerological stations.

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The data sets used and the technique applied are described in Section 2. The general results for the time-averaged budget of zonal momentum over North America are presented in Section 3. A discussion of the residual force in relation to subgrid-scale processes is presented in Section 4, followed by additional comments on the results and their implications in Section 5.

2. Technique and data sets used

For time-averaged (steady-state) conditions, the momentum equation for zonal motion in a spherical pressure-coordinate system can be written as

$$0 = \bar{A} + \bar{B} + \bar{R}_u, \tag{1}$$

where

$$\bar{A} = -\nabla\cdot\overline{u\mathbf{V}} + \frac{\overline{uv}}{a}\tan\phi$$

$$= -\nabla\cdot\bar{u}\bar{\mathbf{V}} - \nabla\cdot\overline{u'\mathbf{V}'} + \frac{\bar{u}\bar{v}}{a}\tan\phi + \frac{\overline{u'v'}}{a}\tan\phi, \tag{2}$$

$$\bar{B} = f\bar{v} - \frac{1}{a\cos\phi}\frac{\partial\bar{\Phi}}{\partial\lambda} = f(\bar{v} - \bar{v}_g) = f\bar{v}_a, \tag{3}$$

$$\bar{R}_u = -\frac{\partial}{\partial p}\overline{u\omega} + \bar{F}_u$$

$$= -\frac{\partial}{\partial p}\bar{u}\bar{\omega} - \frac{\partial}{\partial p}\overline{u'\omega'} + \bar{F}_u. \tag{4}$$

In these equations the overbar denotes a time average, the prime a departure from the time average, a is the radius of the earth, λ longitude, ϕ latitude, p pressure and f the Coriolis parameter. The horizontal wind vector $\mathbf{V}$ ($=u\mathbf{i} + v\mathbf{j}$), the vertical velocity ω ($=dp/dt$) and the geopotential Φ refer to the large-scale flow with horizontal scale larger than a few hundred kilometers; the subscripts g and a refer to the geostrophic and ageostrophic components, respectively. $\bar{A}$ represents the mean horizontal convergence of zonal momentum due to both stationary and transient motions, and $\bar{B}$ the sum of the mean Coriolis force and pressure gradient force in the zonal direction. Both $\bar{A}$ and $\bar{B}$ can be evaluated from the available data. $\bar{R}_u$ denotes the mean residual zonal force needed for balance. It equals the sum of the convergence of vertical transport of zonal momentum by large-scale motions $[-\partial(\overline{u\omega})/\partial p]$, and the zonal friction force due to subgrid-scale processes ($\bar{F}_u$). The latter term represents the effects of processes with scales ranging from the subsynoptic to the molecular. It should be emphasized that when the terms $\bar{A}$ and $\bar{B}$ are evaluated from observations, the residual $\bar{R}_u$ also includes possible systematic errors. These errors could originate from systematic biases in the observing systems (e.g., the selective loss of reports in cases of strong winds, and the possible unrepresentativeness of observations in regions of high terrain), and in the data assimilation procedures (e.g., the influence of the first guess field in the analysis process). Such data deficiencies as random erroneous observations, which are not detected in the data checking routines, are probably non-systematic in nature, and hence would not contribute significantly to the uncertainties associated with the long-term average of $\bar{R}_u$.

Two data sets were used for the estimation of the terms $\bar{A}$ and $\bar{B}$. The first one was prepared at the Geophysical Fluid Dynamics Laboratory (hereafter referred to as GFDL data). It covers the 5-year period from May 1968–April 1973 and is part of a larger data library based on time series of observations at aerological stations. At each station monthly circulation statistics were evaluated, and global horizontal analyses were then made using the method described by Oort and Rasmusson (1971). The analyses for the wind field and the geopotential height field were performed independently, so that no geostrophic constraints were imposed. The GFDL data were available for different seasons.

The second data set was prepared by Lau (1978) using twice-daily hemispheric analyses produced by the National Meteorological Center (hereafter referred to as NMC data). It covers the 11 winters from 1965/66–1975/76. The wind data for the 1969/70 winter were missing. At a given synoptic time, the 12 h forecast is used as the first-guess field. However, the final analyzed field for each variable over regions of good data coverage was basically determined by the aerological observations available for that variable. These are important considerations because a reliable estimation of the ageostrophic velocity requires independent estimates of wind and horizontal pressure gradient.

The rationale for employing the two independent data sets in this study is that we intend to establish the credibility of our estimates for various terms in (1). This is achieved by demonstrating the similarities in the results derived from the two independent analysis procedures. We shall hence emphasize the comparison of the estimates from the two data sets in the following discussions.

Since the GFDL wind data below 850 mb are generally poor, and the NMC wind data are missing at those levels, no results will be shown below 850 mb. At levels above 200 mb, the sample size for the GFDL wind data is slightly smaller than that for the corresponding geopotential height data, so the results for levels above 200 mb will also not be shown.

The area of investigation was chosen *a priori* such

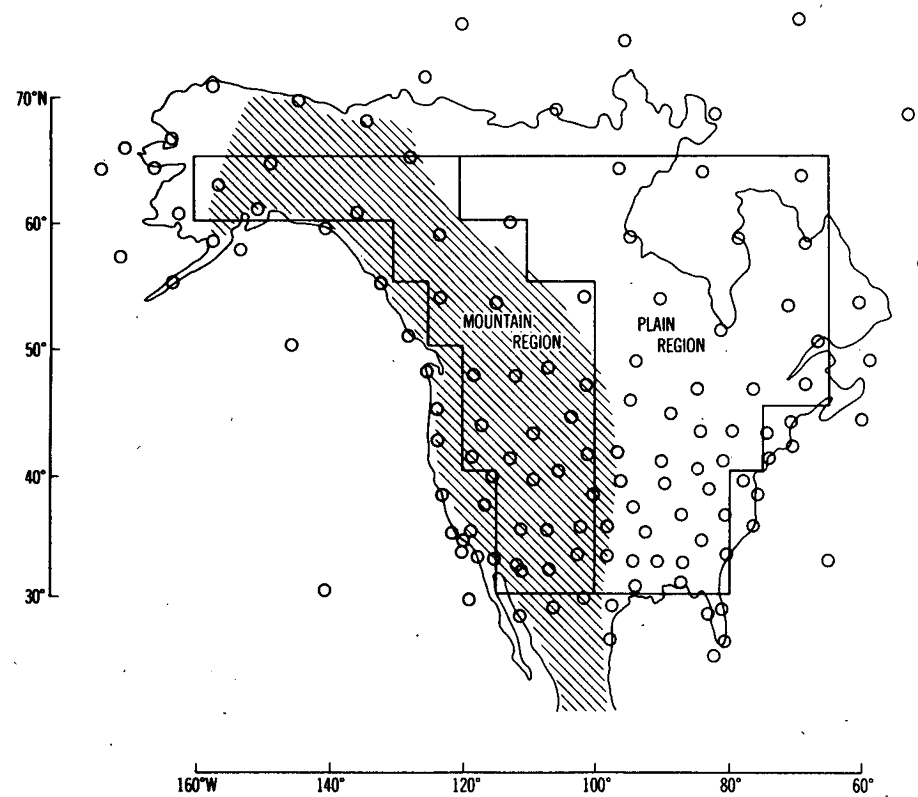

FIG. 1. Distribution of rawinsonde stations in the two regions considered in the analysis of the time-mean budget of zonal momentum. The shaded region indicates local mean topographic heights > 500 m.

that it lies within a network of good data coverage (see Fig. 1). The basic calculations were made for the winter season at each pressure level (between 850 and 200 mb) for "boxes" with sides of 2.5° latitude and 5° longitude. The areal averages formed from the box values for the eastern "plain region" and the western "mountain region" will be discussed.[2] Since the aerological observations in the mountain region may not be representative of the synoptic-scale flow due to the local terrain effects, more emphasis will be placed on the discussion of the results for the plain region.

3. Budget of zonal momentum over North America

The wintertime estimates of terms $\bar{A}$, $\bar{B}$ and $\bar{R}_u$ in (1) from the two data sets are shown in Figs. 2a and 2b for eastern and western North America, respectively. The fact that both data sets give similar results makes these estimates more reliable.

The results for eastern North America (Fig. 2a) show that a mean thermally direct circulation prevails, with poleward ageostrophic flow in the upper troposphere, thus implying a production of zonal momentum at these levels ($\bar{B} > 0$). In a case study of a jet stream system over North America, Mahlman (1973) has also shown that when the flow normal to the jet stream axis is averaged along its length, a thermally direct mean transverse circulation is obtained. This production is in part counterbalanced by a net outflux of zonal momentum ($\bar{A} < 0$). A similar picture of the momentum balance in the jet entry regions was obtained in more qualitative terms by Blackmon *et al.* (1977), whereas the reverse situation prevails in the jet exit region over the British Isles (see H). In the upper troposphere, a relatively large decelerating force ($\bar{R}_u < 0$) is needed for balance, with maximum amplitude at 300 mb. However, between 500 and 850 mb, there is need for an accelerating effect on the mean zonal flow ($\bar{R}_u > 0$). At 700 mb $\bar{R}_u$ is seen to be a term of primary importance, since $\bar{A}$ and $\bar{B}$ have the same sign at this level.

Fig. 2b shows the results of the zonal momentum budget for the western, mountainous part of North America. The results are shown only for levels above 700 mb. The ageostrophic meridional motion over this region also produces zonal momentum through the Coriolis force ($\bar{B} > 0$). However, the

[2] In the averaging process, the box values are weighted according to the cosine of latitude.

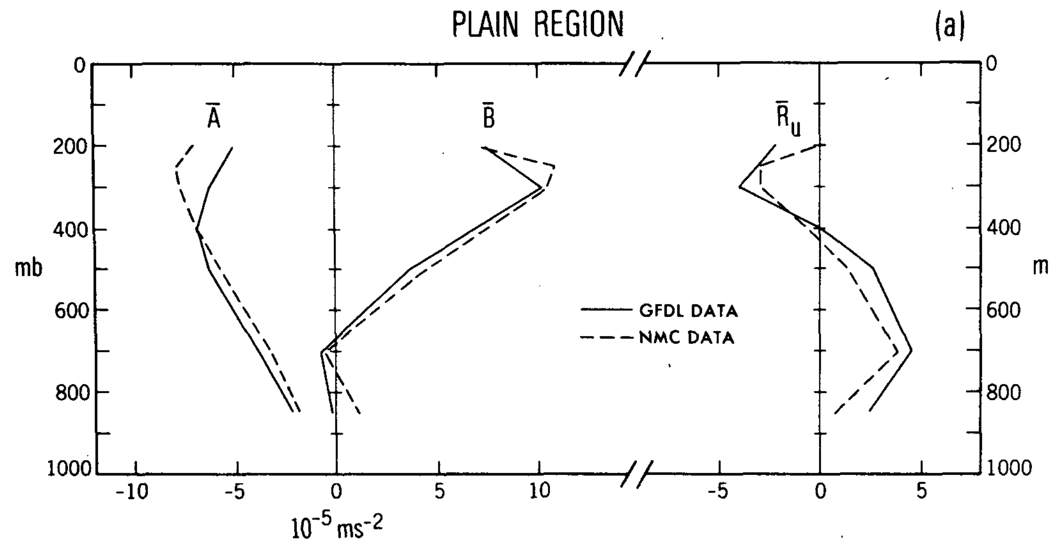

FIG. 2a. Wintertime budget of zonal momentum over the plain region of North America as obtained from GFDL data (solid line) and from NMC data (dashed line). Units: 10^{-5} m s^{-2}.

$$\bar{A} = -\nabla\cdot\overline{u\mathbf{V}} + (\overline{uv}/a)\tan\phi, \quad \bar{B} = f(\bar{v} - \bar{v}_g), \quad \bar{R}_u = -\partial(\overline{u\omega})/\partial p + \bar{F}_u.$$

maximum production occurs at 700 and 200 mb, rather than at 300 mb. A net outflux of zonal momentum also takes place over the mountain region ($\bar{A} < 0$). Compared with the corresponding values for the plain region, it is much smaller and more concentrated in the upper troposphere. The residual term $\bar{R}_u$ in Fig. 2b has a vertical profile very different from that in Fig. 2a. At the levels below about 400 mb a decelerating force ($\bar{R}_u < 0$) is obtained with largest amplitude at 700 mb. The results for $\bar{R}_u$ (and also for $\bar{A}$ and $\bar{B}$) from the two data sets, even though they are qualitatively similar, deviate much more from each other in Fig. 2b than in Fig. 2a. While examining the differences between the results derived from the two data sets, one should keep in mind that these data sets were compiled for different time periods.

In order to examine how the term $\bar{A}$ is partitioned into contributions from stationary $[-\nabla\cdot\bar{u}\bar{\mathbf{V}} + (\bar{u}\bar{v}/a)\tan\phi]$ and large-scale transient $[-\nabla\cdot\overline{u'\mathbf{V}'} + (\overline{u'v'}/a)\tan\phi]$ motions, the areal averages of these individual contributions over the two regions are displayed in Table 1. At 300 mb, where the amplitude of transient fluctuations reaches a maximum, the contribution of the large-scale transient disturbances over the plain region is weaker than that

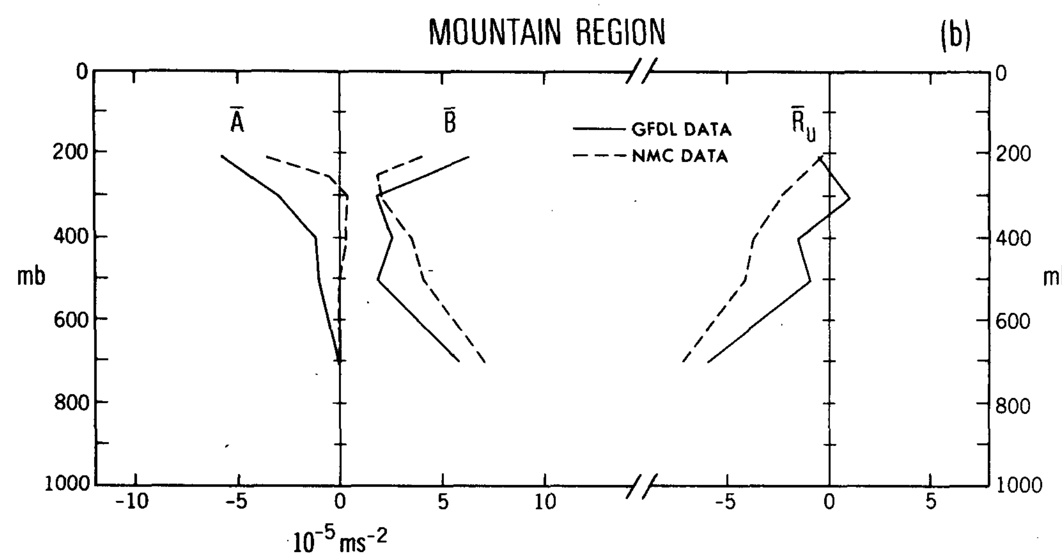

FIG. 2b. As in Fig. 2a except for the mountain region of North America.

TABLE 1. Areal averages of the contributions to the term $\bar{A}$ [see Eq. (2)] by stationary and transient motions during winter, based on the GFDL (NMC) data sets. Units: 10^{-5} m s^{-2}.

	Level (mb)	$-\nabla\cdot\bar{u}\bar{\mathbf{V}} + a^{-1}\bar{u}\bar{v} \times \tan\phi$	$-\nabla\cdot\overline{u'\mathbf{V}'} + a^{-1}\overline{u'v'} \times \tan\phi$	$\bar{A}$
Plain region	200	−4.8 (−6.8)	−0.3 (−0.2)	−5.1 (−7.1)
	300	−4.9 (−6.3)	−1.3 (−1.4)	−6.2 (−7.7)
Mountain region	200	−10.3 (−9.1)	4.5 (5.3)	−5.8 (−3.9)
	300	−8.4 (−6.5)	5.5 (6.9)	−2.9 (0.4)

of the stationary flow by a factor of 3–4. The effect of the transient eddies becomes more significant over the mountain region, where the contributions of the time-mean and transient motions tend to oppose each other.

4. Residual force in the zonal direction

In this section we will examine the residual force $\bar{R}_u$. Values obtained by using the residual technique must always be considered with caution, because spurious (non-physical) values can easily arise due to data deficiencies, sampling errors and numerical techniques used in the calculations.

An obvious question to ask is to what extent are the values of $\bar{R}_u$ shown in Figs. 2a and 2b significantly different from zero in light of the variation of this term from one box to another within the two regions. This can be judged from Figs. 3a and 3b, which give the local values of $\bar{R}_u$ for 700 and 300 mb, respectively.[3] It is seen that at 700 mb a vast majority of the values in the eastern (western) part are positive (negative). It should be noted that the boundary between the two areas was determined on the basis of topography (see Fig. 1) before knowing the results for $\bar{R}_u$. At 300 mb the values tend to be negative in most places, although they are more variable than at 700 mb.

[3] For the sake of clarity, the local values of $\bar{R}_u$ for boxes with sides of 5° latitude and longitude are displayed in Fig. 3. As was mentioned in Section 2, the basic computations were performed on a 2.5° (latitude) × 5° (longitude) grid.

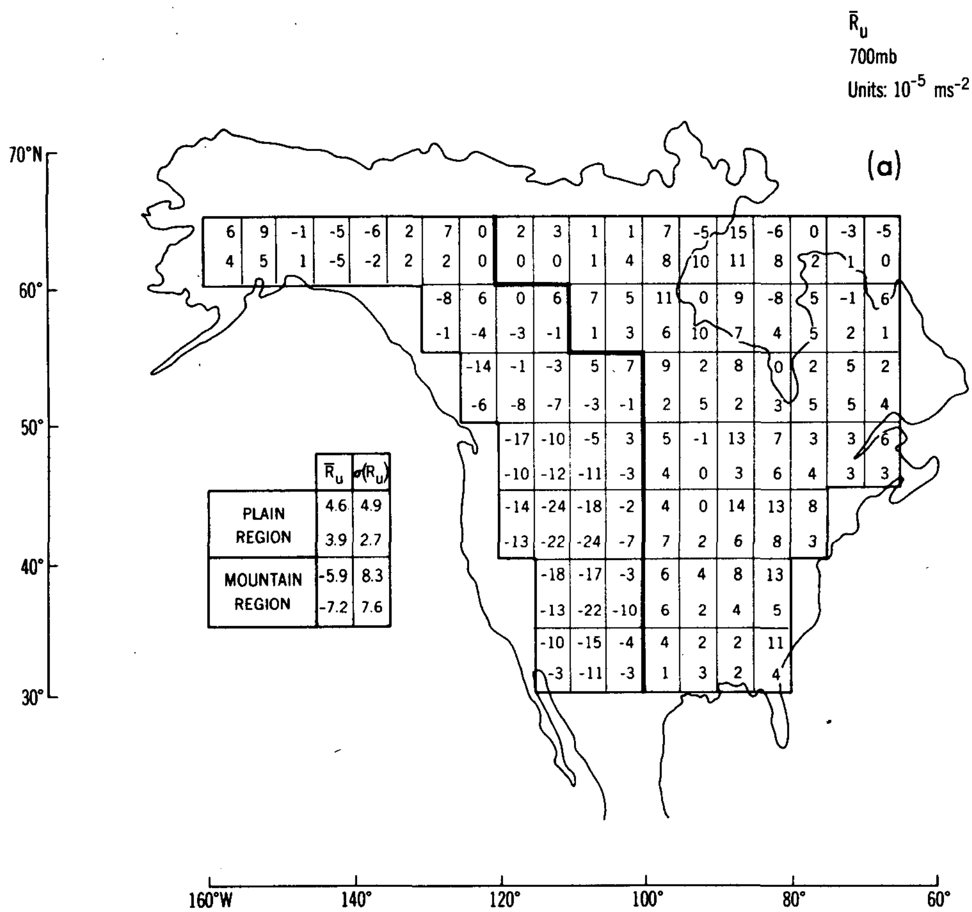

FIG. 3a. Residual force $\bar{R}_u$ in winter for the different grid boxes at 700 mb. Means and standard deviations of $\bar{R}_u$ for the plain and mountain regions are shown in the lower left corner. The upper (lower) numbers are based on GFDL (NMC) data. Units: 10^{-5} m s^{-2}.

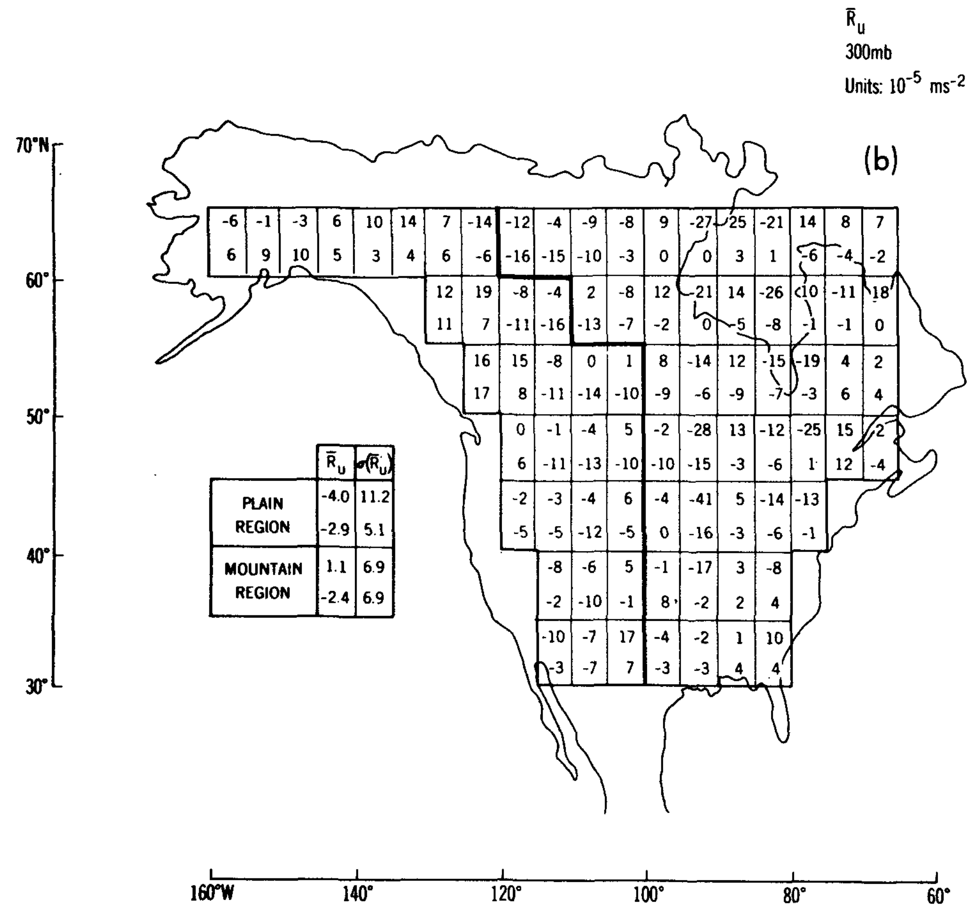

FIG. 3b. As in Fig. 3a except for 300 mb.

As a measure of the variability of $\bar{R}_u$ within a given region, the standard deviations of this quantity $[\sigma(\bar{R}_u)]$ are also tabulated in Figs. 3a and 3b. Making the crude assumption that the values in various boxes are mutually independent, the standard error of the mean $[\sigma(\bar{R}_u)/\sqrt{N}$, where N is the number of boxes in the region being considered] can be obtained by multiplying $\sigma(\bar{R}_u)$ for the plain and mountain regions by 0.15 and 0.18, respectively.[4] On this basis the residual forces are found to be significantly different from zero, except for the mountain region at 300 mb. It is also worth noting that the box values based on the GFDL data exhibit a greater variability than the NMC results.

The results for $\bar{R}_u$ over the plains region are shown for different seasons in Fig. 4 based on the GFDL data. The profiles are qualitatively similar for all seasons. The positive values in the lower troposphere are largest in winter and fall.

We now consider in more detail the expression for $\bar{R}_u$ in (4). The systematic errors associated with data uncertainties will be neglected. Some support for this comes from the similarity of the results for the two data sets. Even though aero-

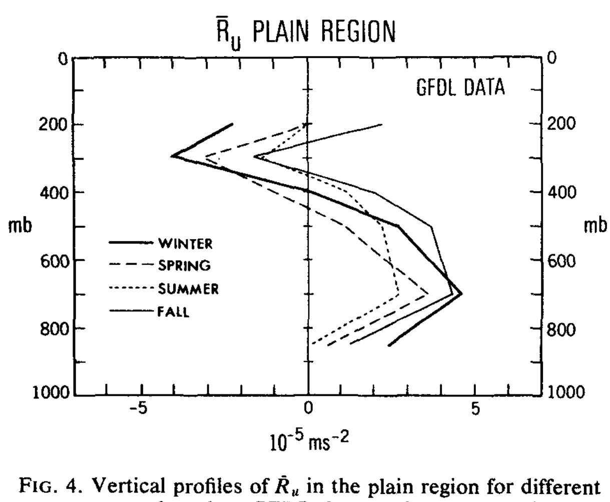

FIG. 4. Vertical profiles of $\bar{R}_u$ in the plain region for different seasons, based on GFDL data. Units: 10^{-5} m s^{-2}.

[4] Since the box values in Figs. 3a and 3b appear to exhibit a certain degree of spatial coherence, the values for neighboring boxes are strictly not independent of each other, and the actual number of degrees of freedom is less than N.

logical observations are the basic material for both data sets, the analysis procedures used are completely independent and very different. If there were some spurious patterns due to the characteristics of the analysis systems, they would probably show up as a difference in the residual $\bar{R}_u$ as computed from the two data sets.

No accurate estimation of the term $-\partial(\overline{u\omega})/\partial p$ in (4) is possible. However, a rough order-of-magnitude estimate can be obtained using the 6 h forecast fields of vertical velocity produced by the NMC models at 500 mb. A more detailed description of these vertical velocity data is given by Lau (1979). A typical value of the terms $\bar{u}\,\bar{\omega}$ and $\overline{u'\omega'}$ at 500 mb over North America was found to be 2×10^{-3} m mb s^{-2}. When the boundary conditions that $\overline{u\omega} = 0$ at 1000 and 0 mb are used, this gives an estimated value of 4×10^{-6} m s^{-2} for the magnitude of $-\partial(\overline{u\omega})/\partial p$ in the lower and upper troposphere. This is one order of magnitude smaller than the corresponding values of $\bar{R}_u$. We can thus conclude that the large-scale vertical flux convergence of zonal momentum cannot account for the relatively large values of $\bar{R}_u$ in Fig. 2.

On the basis of the above considerations, $\bar{R}_u$ should represent mainly the effects of subgrid-scale processes, i.e.,

$$\bar{R}_u \approx \bar{F}_u = \bar{F}_u^H + \bar{F}_u^V, \qquad (5)$$

where $\bar{F}_u^H$ and $\bar{F}_u^V$ are the mean forces in the zonal direction due to subgrid-scale processes taking place in the horizontal and vertical planes, respectively. Having no means of separating quantitatively the values of $\bar{R}_u$ into contributions from $\bar{F}_u^H$ and $\bar{F}_u^V$, the following considerations are necessarily somewhat speculative. In view of the different nature of the processes thought to be responsible for the residual in the lower (below 500 mb) and upper (above 500 mb) troposphere, we will discuss the results for these two layers separately.

a. Lower troposphere

We make the common assumption that subgrid-scale processes in the vertical plane dominate in the lower troposphere (i.e., $\bar{R}_u \approx \bar{F}_u^V$). The positive values of $\bar{R}_u$ between 850 and 500 mb over the plain region (Figs. 2a and 3a) should then be due to a vertical momentum flux into this layer. One possible mechanism for momentum transport from below is associated with the circulation in a cold front. From a synoptic perspective, mesoscale circulations associated with cold fronts in the lower troposphere are usually more intense than in warm front situations. A cold front in the plain region is typically oriented from southwest to northeast. Across such a front there tends to be a positive correlation between the deviations of the zonal and vertical wind components from their respective large-scale background values, which implies an upward flux of zonal momentum by the frontal (subgrid) scale motions in the lower troposphere.

A detailed case study by Browning and Harrold (1970) shows that the band of relatively large upward velocity in a cold front has a width of only 5–10 km even at a height of 1 km. In the framework of an ordinary forecasting model, features of this scale will, for a long time to come, be treated implicitly as subgrid-scale processes.

Going now to the results for the mountain region (Figs. 2b and 3b), a physical explanation of the mean negative $\bar{R}_u$ in deep layers of the troposphere could be that the mesoscale features of the Rockies exert a decelerating effect on the impinging westerly flow through the wave-drag mechanism (e.g., Bretherton, 1969; Lilly and Kennedy, 1973). This forcing effect probably occurs at different levels in different synoptic situations (according to theory, at those levels where wave absorption takes place). The mean climatological effect of mesoscale mountain waves on the large-scale zonal flow could possibly be distributed throughout the troposphere in the way seen in Fig. 2b. Due to the problem mentioned in Section 2 that upper air observations in mountain regions may be unrepresentative for the large-scale flow, the interpretation of the values of $\bar{R}_u$ in Figs. 2b and 3b in terms of momentum flux divergence due to mountain waves should be made with caution.

b. Upper troposphere

Since deep cumulus convection is relatively rare during the winter season, the vertical momentum exchange associated with this process (through $\bar{F}_u^V$) probably does not contribute significantly to $\bar{R}_u$ in the upper troposphere. An alternative mechanism for vertical momentum mixing would be clear air turbulence. However, in order to explain the values of $\bar{R}_u$ at 300 mb in Fig. 2a with an expression of the form $\bar{F}_u^V = K\partial^2\bar{u}/\partial z^2$, a very large value (of order 30 m^2 s^{-1}) for the eddy diffusivity K would be required. Furthermore, this kind of parameterization cannot explain the predominantly positive values of $\bar{R}_u$ over the northwestern part of North America (see Fig. 3b). Therefore, it is reasonable to consider if the values of $\bar{R}_u$ can be explained by horizontal processes (through $\bar{F}_u^H$).

In H, horizontal subgrid-scale processes in synoptic situations characterized by the presence of strong, narrow jet streaks in the upper troposphere were suggested as a partial explanation of the results for $\bar{\mathbf{R}}$ ($=\bar{R}_u\mathbf{i} + \bar{R}_v\mathbf{j}$) over the British Isles. These narrow jet streaks contribute to the momentum budget mainly through that portion of $-(a\cos\phi)^{-1}\partial(\overline{u'u'})/\partial\lambda$ which cannot be resolved by the synoptic rawinsonde network. The effects of such streaks tend to be smoothed out in the NMC analysis procedures, while in principle they should be retained in the station statistics used in the GFDL data set. However, in practice, the contributions of these jet streaks are probably also underestimated in the GFDL data set because they are only rarely detected by the rawinsonde stations. An adequate monitoring of the streaks would require a data set covering a period much longer than the five years used presently. Holopainen and Nurmi (1979, 1980) noted that in a diffluent jet, which occurs frequently over western Europe, the force $\bar{F}_u^H$ tends to accelerate the large-scale flow ($\bar{R}_u > 0$). A similar situation of a diffluent jet often prevails over northwestern North America, where we also find positive value of $\bar{R}_u$ (see Fig. 3b). The present results further suggest that over eastern North America, where a confluent upper air flow predominates, the reverse situation occurs, *viz.*, the force $\bar{F}_u^H$ tends to decelerate the large-scale flow ($\bar{R}_u < 0$).

5. Some final comments

To the authors' knowledge, no comprehensive study of the long-term budget of zonal momentum over North America has so far been reported in the literature. However, Kung and his collaborators (e.g., Kung, 1967; Kung and Baker, 1975) have made a study of the long-term balance of large-scale kinetic energy over North America. They calculated the effects of subgrid-scale motions as a residual term in the kinetic energy equation. One of their findings is that energy "dissipation" occurs both in the atmospheric boundary layer and close to the jet stream level.

Our results for the zonal momentum budget are basically consistent with these studies of the kinetic energy balance. Since the mean budget of kinetic energy over regions of prevalent westerlies is largely determined by forces acting in the zonal direction, our results for $\bar{R}_u$ can be used to estimate the dissipation rate. From Figs. 2a and 2b, a mean value of $\bar{R}_u$ over North America at 300 mb is found to be -2.7×10^{-5} m s^{-2} (NMC data). When multiplied by a typical zonal wind speed of 20 m s^{-1}, we obtain an estimate of the dissipation rate of -5.4×10^{-4} m^2 s^{-3}, which is close to the value (-0.5 W m^{-2} for the 350–250 mb layer, which is equivalent to -5×10^{-4} m^2 s^{-3}) reported by Kung and Baker (1975, Table 8).

If our interpretation of the relative role of $\bar{F}_u^H$ and $\bar{F}_u^V$ in the results for the residual force is correct, the secondary maximum of dissipation found by Kung and collaborators at the jet stream level need not be entirely related to vertical mixing processes such as clear air turbulence (Trout and Panofsky, 1969). It could be mainly caused by horizontal subgrid-scale processes (see also Holopainen and Eerola, 1979).

An important question naturally concerns the relevance and implications of the results reported here (and in H) for numerical forecast models, where presently a horizontal grid size approaching 100 km (or an equivalent truncation wavenumber in spectral models) is often used. Are the residual forces obtained by the technique used here just uninteresting "errors," which arise due to the fact that the data network cannot describe the flow pattern accurately enough, and would they disappear when we have a denser network?

To address the above issues we must clearly distinguish between the computational grid size L_{comp} and the data grid size L_{data}. L_{comp} can be reduced when the speed and capacity of computers increase. However, for the large-scale processes, L_{data} will remain several hundred kilometers for a long time to come (JOC, 1973).[5] When a forecast model, with $L_{\text{comp}} < L_{\text{data}}$, is run with the initial conditions describing only scales larger than L_{data}, the forcing from scales smaller than L_{data} should be included in the model initially. In the course of the integration, the model will develop its own finestructure at scales $L_{\text{data}} > L > L_{\text{comp}}$. It is not obvious that this is treated properly in most of the present models. Since interactions among different scales may have important implications for the predictability of the large-scale flow (Lorenz, 1969), diagnostic investigations of the interaction between the large-scale flow and scales smaller than L_{data} are much needed.

A natural extension of the work reported here would be to investigate the momentum balance below 850 mb with the aid of data more suitable than those used here. Also, calculations should be made of the budget of zonal momentum for various synoptic situations characterized by a different degree of frontal and convective activity. If mesoscale frontal circulations really are, as hypothesized above, one of the basic reasons for the positive mean values of $\bar{R}_u$ in the lower troposphere over eastern North America, a classification of daily budgets of zonal momentum could be made on the basis of a suitable index for frontal or convective activity. This would clearly demonstrate the difference between frontal and nonfrontal cases. However, such a classification cannot be made on the basis of the circulation statistics used here.

The results obtained here using a residual technique can by themselves not be convincingly related to a particular physical mechanism. There-

[5] JOC (Joint Organizing Committee), 1973: The first GARP global experiment: Objectives and plans. GARP publication series, No. 11, 107 pp.

fore, there is a need for a more direct approach. For example, the availability of data from Doppler radar observations and mesoscale field experiments offers an excellent opportunity for a detailed description of both the large-scale and the subgrid-scale motion fields. Appropriate diagnostic studies based on such data sets should yield new insights into the interactive processes between different scales.

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