The Structure of Mean Circulations

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Abstract

A discussion of the best mode of expressing the facts of the mean westerly flow in temperate latitudes parametrically leads to the result that over North America the shape of the vertical geostrophic westerly wind profile is nearly invariant with respect to time and place. The vertical wind profile can be expressed in terms either of the change pressure-gradient with height, or of contour-height gradient with pressure-level. In either case the profile shape is characteristic, and varies little with season, longitude or latitude over North America.

The intensity of the mean circulation varies rythmically with season, but that for any one month may vary markedly from year to year.

The profile shape for an oceanic region differs from that of a land region such as eastern North America in exibiting a lower ratio of high-level to surface wind, and in being much more variable from month to month.

It is shown that the pressure-difference at 7 km between two latitudes, or the difference in mean height of the 200 mb surface gives a measure of the total geostrophic momentum between those latitudes. The square of these quantities also gives a measure of the total kinetic energy of the westerlies between these two latitudes.

1. Introduction

The accumulating mass of aerological material over the past ten years has permitted the construction of mean meridional cross-sections of geostrophic westerly winds. This is a graphic method of conveying the salient characteristics of the mean circulation pattern in some areas of the globe, but it should not be regarded as the ultimate in descriptive aerology because it happens to be the mode best adapted for textbook presentation.

The present paper will be devoted to an alternative, parametric representation of general geostrophic westerly flow in selected areas. The scientific advantage of parametric representation is that it describes mean flow in quantitative terms. This is obviously necessary if we are ever to have a quantitative theory of the general circulation, for a quantitative theory requires quantitative facts to fit it.

The two methods are essentially alternative presentations of the same facts. From a complete parametric presentations, one may readily derive a corresponding cross-section, and alternatively, a complete set of parameters can be inferred from an accurate cross-section.

We shall find, in passing, that an investigation of parameters apt for describing mean flows leads to considerably simplified techniques which may be employed in deriving cross-sections.

2. Mean geostrophic circulation as a function of latitude

If we plot the mean pressure at a fixed height at points along a meridian against the cosine of the latitude, we get a curve the slope of which is at each point proportional to the Westerly geostrophic momentum per unit volume, for, if

$$p = f(\cos \varphi),$$

$$\varrho U_g = \frac{-1}{2\Omega a \sin\varphi} \frac{\partial p}{\partial \varphi} = \frac{1}{2\Omega a} \frac{\partial p}{\partial \cos\varphi}$$

where Ω is the earth's rotational speed, a the earth's radius, ϱU_g the density of westerly geostrophic momentum.

When ϱ, the density, is known, the value of U_g, the geostrophic wind-speed may be inferred.

HESS (1948) has published summer and winter mean ascents for stations close to meridian 80° W, ranging from Arctic Bay, latitude 73° 16′ N, to Salinas, 2° 12′ S. From these data the winter pressure-profiles at heights 3 km (lower, curve) and 10 km, are plotted as a function of cos φ in fig. 1.

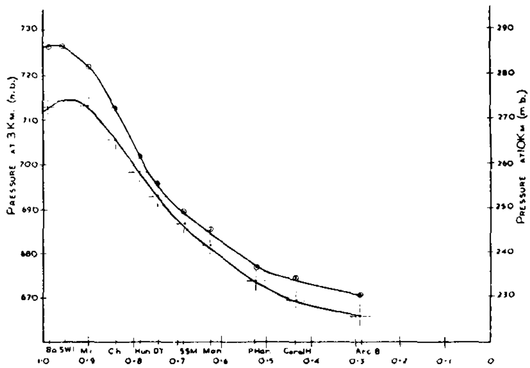

Fig. 1. Pressure at 3 k. m. (starred points) and at 10 km (circles) as a function of Cos φ, meridian 80° W. approx., winter.

It will be noticed that along much of their length, say between the latitudes of Detroit and Monsonee (42° and 56° respectively), the two curves are almost parallel. We find the same shape of curve at other levels between 3 and 10 km. Below 3 km the profile slope falls off. We infer from this parallelism between about 40° and 50° that *momentum is approximately invariant with height in the middle and upper troposphere in winter over Eastern America.*

At latitudes lower than about 40° the two curves begin to diverge, the slope of the upper curve being the greater. Hence below about latitude 40° in winter momentum *increases* with height in the troposphere.

We remark that the slope of the 3 km curve increases steadily with decreasing latitude up to between Huntingdon and Charleston, say in latitude 36°. This constitutes a latitude of maximum momentum and is termed the *jet latitude*. For lower latitudes still the momentum decreases, and becomes zero at the maximum of the pressure profile curve, in latitude 20° approximately. This point we shall term the *doldrum latitude*.

Examining the 10 km profile we find a main jet latitude, again at 36°, and a subsidiary, weak one between Monsonee and Port Harrison, say at latitude 55°. The doldrum latitude of this level is 14°.

The locus of the doldrum latitudes at all levels is termed the *doldrum axis*. It will be observed that the doldrum axis slopes equatorward with height, being in latitude 30° at the surface, 20° at 3 km, and 14° at 10 km.

The *jet axis* is the locus of the jet latitudes at all levels. The main jet axis is practically vertical in latitude 36°. The secondary jet, in latitude 55° does not reach down to the 3 km level.

As we shall see in the next section there is a tendency for momentum to increase with height up to a certain level, and then to decrease. The height of maximum momentum is termed the *momentum jet height*, and the locus of momentum jet heights for all latitudes may be called the *momentum jet profile*. The momentum jet height is found to be approximatly 7 km. The intersection of the jet axis with the momentum jet profile gives the point at which momentum is a maximum. It is termed the *momentum jet centre*. The coordinates of the main momentum jet centre are $\varphi = 36°$, $z = 7$ km. For the subsidiary jet we have, $\varphi = 55°$, $z = 7$ km.

The momentum jet height is not the same as the *velocity jet height;* on account of the variation of density with height the maximum of velocity is found on the jet axis at a height of about 12 km. In describing the velocity field we define a *velocity jet profile* and a *velocity jet centre* in a way exactly analogous to the corresponding momentum parameters.

In order to characterize the westerly geostrophic flow completely we must finally specify an intensity factor. A number of possibilities exist. We could take the momentum at the main momentum jet centre (or velocity at the velocity jet centre). Alternatively a mean momentum between specified latitudes might be used, the best level to take being the height of the momentum jet profile, 7 km approximately. This second alternative is more in accord with current usage, which accords a unique position to the *zonal index* or pressure difference at fixed level (usually the surface) between latitudes 30° and 60°.

The zonal index at the jet level, 7 km, will be shown later to represent the mean momentum of the westerlies at all levels between the assigned latitudes.

Once the zonal index at 7 km (or the momentum at the momentum jet centre), the momentum jet profile, the jet axes and the doldrum axis have all been specified, together with the variation of momentum with height, the structure of the momentum field is completely specified.

Fig. 2 shows the corresponding profiles at 3 and 10 km for Hess' summer data. The salient feature is the reduction of the zonal index as compared with winter. The pressure-drop between Charleston and Port Harrison at 7 km is 23.5 mb in summer, as compared with 42.1 mb in winter. Mean westerly summer momentum is just over one-half the mean winter momentum.

The doldrum axis has migrated to about 28° latitude, and is almost vertical, while the momentum jet centre is now at latitude 55°.

3. The vertical profile of momentum

It has been shown (JAMES unpublished) that the latitudinal pressure-difference between

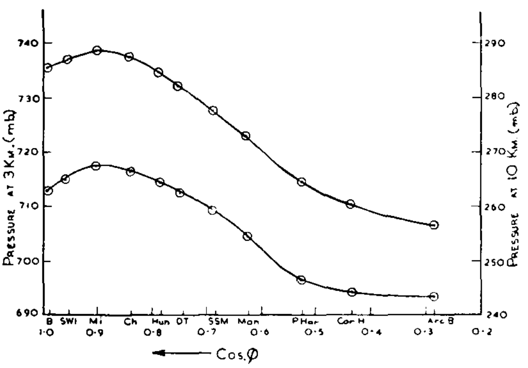

Fig. 2. Meridional summer profiles of pressure at 3 km and 10 km.

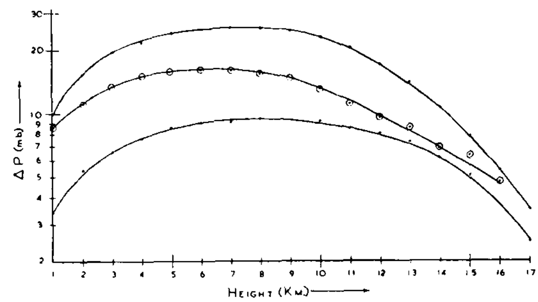

Fig. 3. Pressure at a fixed height between different latitudes. Top curve, Charleston—Sault-Ste. Marie, intermediate curve, Sault-Ste. Marie—Port Harrison, lowest curve, Charleston—Miami.

two points in nearly the same longitude varies with height according to the relation

$$\Delta p = (\Delta p)_m \exp\left[-k(z - Z_m)^2\right] \quad (1)$$

where Z_m is the momentum jet height, and $(\Delta p)_m$ is the pressure difference at this height. We may term $(\Delta p)_m$ the *jet intensity* between the pair of stations concerned.

Fig. 3 shows the relation between Δp and height plotted on a logarithmic scale for station pairs in the Hess winter data. The upper curve gives the pressure difference between Charleston and Sault-Ste-Marie, latitudes 33° and 46½° respectively. The flat maximum occurs at about 7.3 km, the mean momentum jet height for this latitude range, and is practically symmetrical for a range of 5—6 km on either side of this central point. The profile is almost identical with one for Charleston-Caribou as yet unpublished, being closly represented by a profile of the form (1), as may be verified by plotting the cumulated value of Δp against z on probability paper.

The middle curve represents the mean vertical profile of westerly momentum for the latitude range 46½°—58½° (Sault-Ste-Marie and Port Harrison). The mean momentum jet level is at 6.6 km.

The lowest curve, Miami, latitude 26°, and Charleston, is again of the form (1). The momentum jet level is at 8.4 km. The same type of profile is found for the pressure-difference between Port Harrison and Arctic Bay (latitude 73°), the momentum jet height being 6.3 km.

We see, therefore, that the shape of the vertical momentum profile is characteristic over a latitude range, 25° N—73° N of eastern North America. It can be shown that the profile shape (1) holds for West and Central U. S., and in different seasons. It is thus a quasi-invariant of the mean westerly flow over the U.S.

Since the momentum profile shows a very flat maximum the Clayton-Egnell rule, momentum constant with height, gives a good approximation over quite a considerable range of height. For about 2 km on either side of the jet level the reduction in momentum is only about 6 %, while for a range of 8 km centred at the jet level the maximum departure is 20 %.

A parameter specifying the properties of the vertical profile is the *equivalent thickness*, Z_T, defined by

$$Z_T \Delta p_m = \int_0^\infty \Delta p \, dz$$

Z_T is the thickness of an equivalent stratum of air having everywhere a momentum equal to the momentum at jet level, with the same total momentum as the actual atmosphere.

The average winter equivalent thickness is 12.5 km between latitudes 26° and 33°, 11.4 km between 33° and 46½°, 11.5 km between 46½° and 58½°, and 10.0 km for latitudes 58½° to 73°. These are actually underestimates, for the momentum below 1 km and above 16 km. has been disregarded. It can be shown that the equivalent thickness does not vary significantly with season; it is, in fact, a quasi-invariant of the mean circulation.

The total westerly momentum between latitudes φ_1 and φ_2 in a meridional slab of unit thickness extending vertically to infinity is

$$a \int_{\varphi_1}^{\varphi_2} d\varphi \int_0^\infty \varrho U \, dz$$

When one of the latitudes lies in the region of low level easterlies, the lower limit of the z integral becomes the height of the doldrum axis in that latitude.

For the total momentum between, say, latitudes 30° and 60° we may thus write

$$U = \frac{1}{2\,\Omega} \int_{30^\circ}^{60^\circ} \frac{Z_T}{\sin\varphi} \left(\frac{\partial p}{\partial \varphi}\right)_m d\varphi$$

where $\left(\frac{\partial p}{\partial \varphi}\right)_m$ is the pressure gradient along the jet profile. Hence to an approximation, since the equivalent thickness varies only slightly with latitude,

$$U \approx \frac{Z_T\,(\Delta p)_m}{2\,\Omega \sin\bar{\varphi}} \qquad (2)$$

where $\bar{\varphi}$ is some mean latitude.

The total mean westerly momentum between two latitudes is thus proportional to the pressure-difference between 30° and 60° at the height Z_T, or 7 km.

The total momentum over the whole westerlies belt is given by

$$U' = \pi a \, Z_T \operatorname{Cot} \bar{\varphi} \, (\Delta p)_m / \Omega \qquad (3)$$

where $(\Delta p)_m$ must now be interpreted at the zonal index at 7 km between latitudes 30° and 60°.

The zonal index at other levels gives a (more approximate) measure of the total atmospheric momentum, because of the quasi-invariance of the vertical profile-shape. However, the 7 km level is to be preferred as the momentum changes with height least here, and is less subject to random fluctuations. The worst level to take is the surface, as here the profile is subject to fluctuations connected with the boundary conditions which may not be appreciably reflected in the mean flow at height.

4. The use of contours

Hess' data are given in terms of pressure at a fixed level. It is now standard practice in most weather services to quote contour heights of standard pressure surfaces rather than pressures at fixed height levels. Fig. 4 shows the mean contour heights for the year 1946—49, inclusive, as a function of cos φ for U.S. stations around 80° W longitude for the 300 and 700 mb levels.

These profiles show the same salient features as those drawn for pressure at fixed height levels. They have the advantage that the slope determines velocity directly, and not momentum. By reading off the slopes of the contour profiles for a number of different pressure-

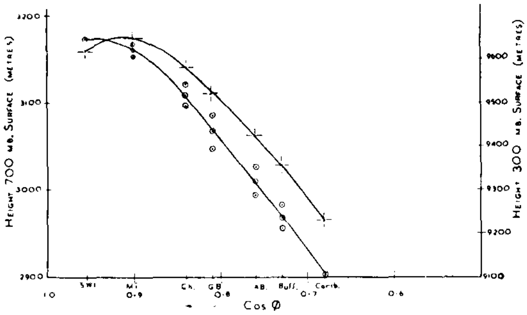

Fig. 4. 700 mb contours (starred) and 300 mb contours (circles) as a function of latitude for meridian 80° W. approx, for 1946—49. The upper circles indicate 300 mb, contour height in the year of greatest height (1949) and the lower circles apply to 1947, the year of least profiles. The intermediate points, through which the smooth curve has been drawn, refer to the mean four year values.

levels, it is possible to compute cross-sections with much greater facility and speed.

Three points (circled) are shown for each station in the 300 mb profile in fig. 4. The middle point is the four year mean contour height, and through these points the smooth curve of the mean profile is drawn. The upper set of points gives the mean contour heights in the year of greatest general contour height (1949), while the lowest set corresponds to 1947, the year of least general contour height. It will be seen that the annual mean height of the 300 mb surface may change by more than 50 metres from year to year, but that the surface rises or falls by much the same amount in all latitudes. If the fall or rise were uniform in all latitudes the slope relations would remain unchanged and the West-wind pattern would be unaltered from year to year.

However, we should expect a greater variability of mean contour height in high than in low latitudes. This is borne out in fig. 4, where it will be seen that the '49 and '47 mean contours differ by 30 m at Miami and by 80 m at Greensboro. We should consequently expect years of high average contour height to be weak jet years, and years of low average height to display strong jets.

The main jet latitude is approximately 36° for both 1947 and 1949, but the westerly speed at the 300 mb level is 33 % higher in 1947 than in 1949. There is a subsidiary maximum between the latitudes of Pittsburg and Buffalo in 1949, and between Buffalo and Caribou, or in higher latitude, in 1947, this subsidiary jet being stronger in 1949. Fig. 4 thus reveals appreciable variations in the strength and structure of the annual mean westerlies from year to year, although the overall pattern is not radically altered. It should be borne in mind, also, that systematic errors, changing from year to year cannot be completely excluded from the estimates of mean 300 mb contour heights.

5. The vertical wind profile

The difference in the mean contour height of a given pressure-surface between two stations is a measure of the mean geostrophic wind component between the stations in a direction transverse to the line joining them. If one station is due North of the other the westerly component of mean wind is measured.

The way in which this height-difference varies with the pressure-surface gives a measure of the change in geostrophic wind-speed with height. In an application such as this it is natural to take pressure as the measure of height. Whether we take as our argument the actual pressure, or its logarithm, as being more nearly proportional to height is purely a matter of convenience. In fig. 5 we plot mean contour height difference between Charleston and Caribou against pressure for the four-year period, 1946—49. The curve shown is the *reduced* profile, normalized so that $(\Delta Z)_{max} = 1$.

It will be noticed that the points lie evenly on a smooth curve which is almost a linear

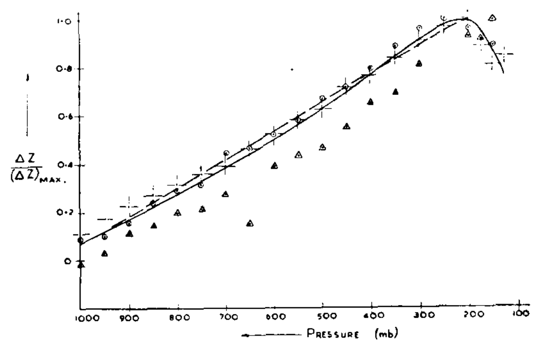

Fig. 5. Mean vertical westerly wind profile between two latitudes. The full curve is drawn through the four year mean contour-differences, Charleston—Caribou. The circled points represent the same relationship for Pittsburg—Caribou, and the starred points are for Greensboro—Pittsburg. The triangles apply to Charleston—Miami.

function of pressure up to jet level. The linear relationship

$$U = U_0 (p - p_m) / (p_0 - p_m) + U_{\max} (p_0 - p) / (p_0 - p_m) \quad (4)$$

where p_0 is the surface pressure and p_m the pressure at jet level, U_0 and U_m being the corresponding westerly geostrophic speeds, is indicated by the dotted line in fig. 5. If a closer relationship is desired a parabolic curve could be fitted, but for many purposes the linear relation will be found quite accurate enough. The jet level is estimated to be 220 mb.

The profile shown is that of mean westerlies over a considerable latitude range. It is therefore of interest to see whether the shape is substantially unchanged for a narrower range of latitude. The starred points in fig. 5 give the vertical profile between Greensboro and Pittsburg, while the circled points show the relationship for Pittsburg-Caribou. The points lie very close to the curve drawn for Charleston-Caribou, so that the profile shape is very nearly independent of the latitude between 33° and 47°. Above 200 mb there is some considerable scatter indicating that observations at the highest levels may not be absolutely representative.

Even below latitude 33° the vertical profile shape is very similar to that for higher latitudes. The points enclosed in triangles give the profile-shape for Miami-Charleston. The scatter is more noticeable, probably due to the small difference in contour height between these stations, but the linear relationship is still a reasonable approximation. The jet level is somewhat higher than in higher latitudes, at 190 mb.

It is also possible to show that the vertical profile-shape does not change significantly with season. In fig. 6 the mean reduced vertical profile, Charleston-Caribou, 1946—49 is again drawn, together with points for the months January, April, July and October, 1949, for the same pair of stations. The points for the first three months cluster very closely around the four-year mean profile, with no significant seasonal change. The most discordant month, October 1949, is indicated by the lower line. The jet level is higher, and the curve more concave than for the annual mean. Even so, the similarity between the two curves is striking.

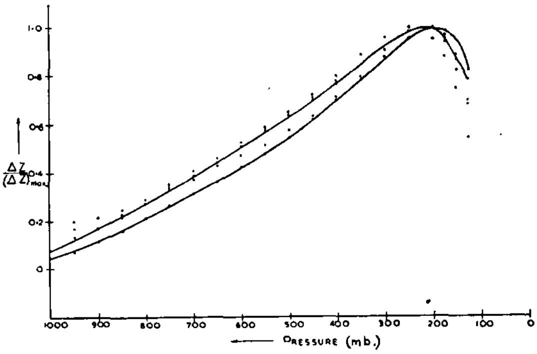

Fig. 6. The four year mean westerly wind profile for Charleston—Caribou — full curve, with indications of seasonal variation. The points refer to the different seasons, but it is apparent that the seasonal scatter of profile shape is slight.

The scatter of monthly values from the four-year mean profile is most pronounced below 850 mb. This is perhaps not unexpected; the wind-field at low levels is most likely to be influenced by surface boundary conditions which may have nothing to do with the general westerly flow-invariance of the wind-speed profile. That it does not vary appreciably with longitude over the U. S. can be inferred from the invariance of its counterpart, the momentum profile.

6. The annual variation of geostrophic wind

We have remarked that the vertical profile of geostrophic mean wind is a quasi-invariant of the atmospheric circulation over the U.S.; it does not change appreciably with season. Hence the seasonal variation of wind at any one pressure-level may be taken as typical

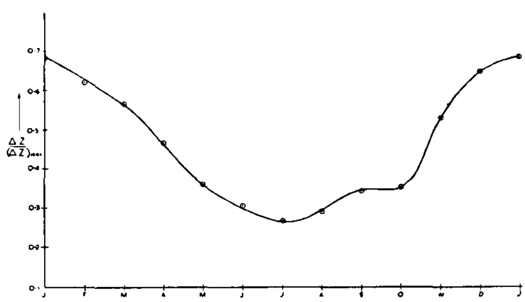

Fig. 7. The seasonal variation of mean westerly wind between Charleston—Caribou.

of the variation at all levels, except, possibly, near the surface.

In fig. 7 we show the mean annual variation of contour-height difference between Charleston and Caribou, 1946—49 at the 200 mb. level.

The geostrophic Westerly speed of the mean circulation reaches a maximum in January and a minimum in July. The curve is approximately sinoidal in shape.

The curve shown is based on four years of data; the periodic curve in any one year may depart markedly from this pattern. Thus the maximum in a given year may well occur in December or February, or even March; the minimum might be displaced to June or August. The strength of the mean monthly circulation in any month may be displaced a month or two from the seasonal mean. In the most variable month, the departure from the seasonal mean will be 25 % or more above or below normal in 30 % of months. Nevertheless, in spite of large 'casual' variations in the strength of the circulation, there is a strong underlying seasonal rhythm. The October plateau, showing up in fig. 7, may be a real effect, or may be due simply to the inclusion of a run of abnormally sluggish Octobers. Such points can only be settled by a longer series of data than that used here.

7. The kinetic energy of middle latitudes

The mean westerly wind-speed at a given pressure level between two stations on the same meridian separated by a latitude distance Δy is

$$U = \frac{g\,\Delta Z}{2\,\Omega \sin \bar{\varphi}\,\Delta y}$$

where ΔZ is the contour-height difference, and $\bar{\varphi}$ is a mean latitude. The total momentum over the latitude distance Δy of a meridional slab of unit thickness is, then

$$\Delta y \int_0^\infty \varrho\, U dz = \frac{1}{2\,\Omega \sin \bar{\varphi}} \int_0^{p_0} \Delta Z\, dp$$

Now we have seen that ΔZ as a function of pressure (height) is a quasi-invariant of the mean circulation

$$\Delta Z = (\Delta Z)_m\, f(p), \quad f(p_m) = 1$$

We may thus write for the total momentum of the slab

$$U = \frac{(\Delta Z)_m}{2\,\Omega \sin \bar{\varphi}} \int_0^{p_0} f(p)\, dp,$$

and for the total momentum of an entire latitude belt, assuming the functional form of $f(p)$ to be the same in all longitudes,

$$\frac{\pi\, a\, (\Delta Z)_m}{\Omega} \cot \bar{\varphi} \int_0^{p_0} f\, dp$$

We see, therefore, that the total westerly momentum in a given latitude range is proportional to the difference in the mean contour heights at the level of maximum geostrophic wind-speed. The integral

$$\int_0^{p_0} f\, dp$$

has the dimensions of pressure. We term it p_T the *pressure-thickness.* It represents the weight, in pressure units of atmosphere, which, possessing everywhere the maximum wind-speed, has the same total momentum as the actual atmosphere.

From the vertical profile of geostrophic speed, already determined, we find,

$$p_T = 540 \text{ mb.}$$

Since f is a quasi-invariant of the westerly circulation, so is p_T. Hence a convenient overall index of the strength of the westerly circulation is the mean contour height difference at a level of approximately 200 mb.

This formulation is in every way equivalent to that in terms of the zonal index, for which we have the equivalent thickness as the quasi-invariant. Whether the pressure thickness or the equivalent thickness is used to characterize the mean vertical profile depends entirely on whether we are dealing with contours or fixed level charts.

The total kinetic energy in a vertical longitudinal slab of unit thickness between two latitudes is

$$K = \frac{1}{2} \int_{\varphi_1}^{\varphi_2} a\, d\varphi \int_0^\infty \varrho\, v^2\, dz$$

Hence

$$K = \frac{g\,(\Delta Z)_m{}^2}{4\,\Omega^2 \sin^2 \overline{\varphi}} \int_0^{p_0} f^2\, dp \qquad (5)$$

We may thus define a *kinetic pressure-thickness*,

$$p_K = \int_0^{p_0} f^2\, dp$$

Since f is quasi-invariant, so is p_K. From the 4-year mean vertical profile between Charleston and Caribou, we find

$$p_K = 385 \text{ mb.}$$

We see from the expression (5) that the total atmospheric geostrophic kinetic energy between two latitudes is proportional to the square of the contour height difference at about 200 mb between the two latitudes.

8. Other longitudes

It is important to know whether the relationships found apply to the N. American continent only, or generally characterize the entire Westerly belt. YEH (1950) has published a winter cross-section along 76° E, using actual rawin data rather than geostrophic speeds. The general structure is strikingly similar to that found by Hess for Eastern America. Two jets occur around latitudes 28° and 42°. Both these positions are further south than the corresponding jets over America, and jet speeds are about 20 % higher. In latitude 40° the speed of jet level is some 6—8 times that at 900 mb, in agreement with what is found over N. America. From the appearance of the cross-section it would seem that wind-speed is a nearly linear function of decreasing pressure, as over N. America. A profile published by CHAUDHURY (1950) shows approximately the same characteristics.

From this it might appear that the structure of the westerlies is approximately the same in all longitudes. At the moment there are no cross-sections available for Western Europe, a circumstance which might call for some surprise, considering the wealth of data available.[1]

Indications point to the structure of the mean circulation near the zero meridian differing appreciably from that found over N. America, or Asia. Fig. 8 shows the mean contour difference (normalized) between Larkhill (latitude 52°) and Lerwick (60°) for the two year period 1948—49. While in a general way the vertical profile for the meridian of Greenwich resembles that found over N. America in corresponding latitudes, there are perceptible differences. The relation between geostrophic speed and decreasing pressure does not show such a close approach to linearity. Moreover, the jet height is closer to the 300 mb surface than to the 200 mb surface The jet speed for Larkhill-Lerwick is only about three times the 950 mb speed, the corresponding ratio over similar latitudes ni N. America being 6 or more.

Perhaps the most striking feature of the westerly circulation near the Greenwich meridian is its great *variability*, as constrasted with the relative steadiness of the flow over N. America. Monthly mean normalized contour height differences are plotted in fig. 8. They show a very wide scatter about the long-term average profile shape, with no significant seasonal variation. In fig. 6 we indicated the monthly 'scatter' of profile shape over the eastern U. S., and comparing the two figures it will be at once apparent that the scatter is altogether greater in the case of the British figures. The two dotted lines in fig. 8 indicate the envelopes of extreme scatter

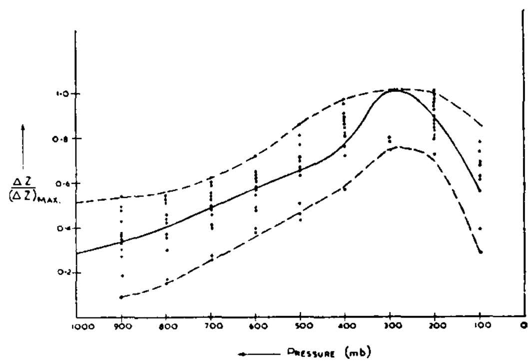

Fig. 8. The mean annual vertical wind profile for the Greenwich meridian (Larkhill—Lerwick). The points indicate the monthly scatter, the dotted lines, the extremes within which the profile-shape may lie.

[1] Since completing this paper the author has drawn a geostrophic cross-section down the Greenwich meridian (Met. Mag. Dec. 1951). This cross-section is in substantial agreement with the conclusions advanced here.

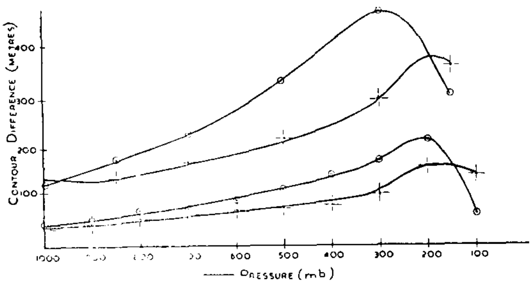

Fig. 9. S. Hemispheric mean vertical west wind profiles. Circles summer, stars winter. Upper set of curves, Macquarie Is.-Laverton, lower set Auckland—Taieri.

about the mean vertical profile of geostrophic wind. It is evident that in any 'extreme' month the profile shape may bear very little resemblance to its long-term mean. We may contrast the profile steadiness over N. America and over U.K. by saying that for 4 months of the year the mean monthly wind at, say, 500 mb will depart by more than 6½ % of the value predicted from the long-term mean profile-shape over N. America, but will depart by 25 % or more from the long-term value in the region of the U.K. In the U. S. it is possible to take the long-term mean profile as a good approximation to the actual monthly profile, but over U. K. this is no longer possible; the profile-shape is not quasi-invariant.

These facts demand further factual investigation. Why is the stability of mean westerly flow so strikingly different in different regions of the earth?

Riehl (1949) has conjectured that the observed stability in the structure of the Westerlies is due to lee effects of high geographical barriers. The westerly flow is notably uniform in geographical barriers. The westerly flow is notably uniform in lee of the Andes (Riehl) and of the Himalayas (Jeh). The quasi-invariance of the westerly flow-pattern over eastern N. America may similarly be attributed to the effects of the Rockies. This explanation would attribute the instability of the flow near the Greenwich meridian to the absence of a high barrier to westward, with a resulting absence of a quasi-stationary lee flow-pattern.

If this explanation were the correct one vertical profiles in the Southern Hemisphere, except in lee of the Andes, may be expected to display a structure similar to that found for the British Stations. In fig. 9 four southern hemisphere profiles are plotted. The top pair relate to summer and winter for the station-pair, Laverton (38° S, 145° E) and Macquarie Island (54° S, 159° E) (Loewe and Radok, 1950). The summer profile is indicated by the circled points. These profiles are more reminiscent of the U. K. than the U.S. profiles. The jet height is nearer the 300 mb level in summer, and the 200 mb level in winter. The ratio of jet speed to 950 mb speed is over 3 in summer and under 3 in winter, in agreement with the value of this ratio observed for Larkhill-Lerwick. One feature not observed in Northern Hemisphere data is that the geostrophic speed at jet level is higher in summer than in winter. This feature is reproduced also in the profiles for Auckland (37° S, 175° E) and Taieri (46° S, 170° E) (Hutchings 1950). The jet height for these two stations is closer to the 200 mb level both in summer and in winter. It will be borne in mind that the Southern profiles are based on very limited data — only one month of Macquarie Island data, and two months from Taieri. However, the profiles shows the same characteristics, although the data are limited, and refer to different times. There is a strong presumption for supposing the mean southern profiles to resemble closely those found in corresponding regions of the Northern Hemisphere.

On the Riehl hypothesis one would attribute the N. American type of vertical wind profile to the presence of an orographic barrier, and the U.K. type of profile to the absence of any barrier to westward. The facts presented here are certainly not inconsistent with such a presumption.

On the other hand, an alternative conjecture equally fits the facts. It might be postulated that the pattern of flow observed over N. America is typical of a continental region, while the U. K. type of profile is equally typical of an oceanic area.

Over a continent there is a considerable surface frictional drag, with a resultant conversion of westerly angular-momentum. I. this loss of A. M. is to be made good by high-level advection, it is reasonable to anticipate strong mean westerlies aloft, and a marked gradient of wind with height.

On the other hand, over an oceanic region the A. M. depletion at the surface is less for a given pressure gradient, and a smaller turn-round of westerly momentum at height will be required to make good the losses. Hence a smaller increase in westerly wind with height will suffice to maintain a steady state.

Of great interest is the apparent unsteadiness of the oceanic circulation, as manifested by the variability of the monthly mean profiles near the Greenwich meridian. A more intensive study of oceanic profiles is necessary if we are to establish the nature of this instability, and its implications for the global balance.

The unsteadiness of the mean profile over Oceanic regions may be a reflection of the actual variability of wind, associated with the frequency of migrating disturbances. BROOKS and CO-WORKERS (1950) have published world maps of standard vector wind deviation for levels between 700 mb and 130 mb. In these charts the global pattern is clear, a maximum variability over the oceans in middle latitudes, and minima over the continents.

Minima are found, at all levels throughout the year, to be associated with the Himalayas, Andes and Rockies. The African minimum shows a decided preference for the highest ground from Tanganyika to Abyssinia. A trough connected with the European Alps is also in evidence up to 500 mb.

This might be taken as evidence favouring Riehl's hypothesis, although it may be difficult to show why the presence of perturbations in the mean flow should affect the stability of the flow itself. Nevertheless, there seems to be a distinct association between wind variability and the year to year variability in mean flow.

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