Some results of the surface wind and atmospheric pressure fields analysis near the Equator

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ABSTRACT

Based on meteorological observations from five USSR research ships during Expedition TROPEX-72 in the Atlantic Ocean for the period from June to August 1972, the relationship between surface winds and pressure gradients near the Equator is analysed. For the narrow Equatorial zone (0–2.5° N) the nearly complete independence of the surface wind speed (rather strong 7–8 m/sec) and direction from the local pressure gradients was found and on this basis the conclusion is drawn concerning the important role of the air stream inertia in this zone. However, the estimates of different terms in the equation of motion for the near-Equatorial latitudes showed that, as in the middle latitudes, the inertia terms of the mean motion are here one order of magnitude smaller than the terms of pressure gradient. Having this in mind we believe that the only proposition which could be made concerns the important role of the inertia associated with turbulent rather than average motion in the dynamics of the Equatorial atmosphere. It is shown that the inertia of the air stream near the Equator can account for the asymmetry of the vorticity field around the axis of the Equatorial trough (cyclonic relative vorticity north of the trough axis and the anticyclonic one south of the axis in Northern Hemisphere). It is shown that the wind deflection angles become normal for the usual geostrophic relation over the sea only north of 11–13° N.

1. Introduction

Up to this day the relations between wind and pressure gradient near the Equator have remained unclear. This is easily explained by shortage of reliable data, unaffected by surface topography wind and pressure, for this region.

In June–August 1972, five research ships of the USSR expedition TROPEX-72 worked in the tropical Atlantic (Petrossiants, 1974). Their observations provided our first opportunity for studying the question in detail.

This paper presents the results of computation of the atmospheric pressure gradients on the basis of TROPEX-72 data and discusses the character of wind/pressure gradient relation for various distances from the Equator.

2. Data, synoptical conditions and pressure gradient computation

The TROPEX-72 data used for our work were surface wind and atmospheric pressure observations collected by five ships at three sites (Table 1). Meteorological observations from the ships were made 8 times a day from 11 July to 12 August and hourly from 13 August to the end of the experiment. All the surface pressure data were corrected for the systematic differences between the five ships' aneroids found through intercomparison at the beginning, in the middle and at the end of the work.

One can judge the synoptic situation in the region under study from Fig. 1 (taken from Peskov et al. 1974). It shows the distribution of atmospheric pressure at sea level and the mean resultant wind on board, averaged over the period of the first synoptic polygon (6–21 August 1972). The axis of the Equatorial trough coincided with the main branch of the intertropical convergence zone located at that time approximately along the 10° N latitude. North of it, north-east trade winds were observed, which changed the direction for northwest near the trough. From the south,

Table 1. *TROPEX-72 ships' positions and periods of observation on the sites*

Polygons... Observational periods...	Mesometeorological 11–25 July 1972		1st synoptic 6–21 August		2nd synoptic 22–27 August	
	lat.	long.	lat.	long.	lat.	long.
Prof. Vyse	10° N	21° W	12.5° N	22° W	10° N	26° W
Ak. Kurchatov	7.5	21	7.5	22	5	26
Passat	7.5	18.5	7.5	17	5	21
Musson	7.5	23.5	2.5	22	0	26
Prof. Zubov	5	21	7.5	27	5	31

a south-east trade wind approached the trough axis. It came from the southern hemisphere, deviated to the right after crossing the Equator to become south-westerly as it approached the trough axis. In July the axis of the Equatorial trough lay, on average, some 2 or 3° of latitude nearer to the Equator than in August; besides the seasonal variation this axis underwent notable daily fluctuations along the latitude.

We computed the pressure gradients from the formula

$$P(x, y) = ax^2 + bx + c + dy^2 + hy$$

which is the result of parabolic approximation of the atmospheric pressure field at the polygons.

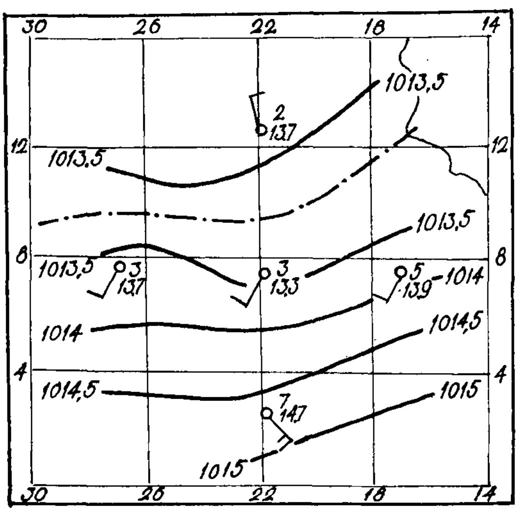

Fig. 1. The mean atmospheric pressure and mean resultant wind distribution at the 1st synoptical polygon (6–21 August 1972) according to Peskov et al. (1974). The dashdotted line is axis of intertropical convergence zone. Upper values by the circles—the atmospheric pressure (P-1000 mb), lower values by the circles—resultant wind speed.

In this way it was possible to obtain pressure gradients for the positions of each of our five ships at the polygons (Table 1) and to compare the values thus obtained with wind measured directly from the ships. In contrast with other methods, we have not lost any information at the extreme north and south points at the polygons.

Pressure gradients were computed from the raw data and also from the data filtered with $f(t) = 1 + \cos 2\pi t/l$ to remove all fluctuations with 12 hour and shorter periods.

The same filter was used for smoothing the surface wind data through separate application of it to zonal and meridional components.

3. Analysis of results

We shall consider first the mean values of the surface wind speed and also atmospheric pressure gradients (Table 2), computed according to the above procedure for the positions of each vessel on the polygons and then averaged over the latitudes for all observation phases (July–August 1972). It can be seen that the pressure gradients here are very small. In just a few cases they were as high as 5 mbar/5° latd., while the average figure was about 1 mbar/5° latd., i.e. 8 times less than in the middle latitudes (Gandin et al., 1955), even though the wind speed for example, on the Equator, was comparable to that in middle latitudes. It is a characteristic fact that the minimal gradients of atmospheric pressure were observed *not on the Equator, but near the axis of the* Equatorial trough zone. The smoothing, as seen from Table 2, practically does not affect the atmospheric pressure gradients, but it strongly affects the values of wind speed decreasing them by a half.

Let us now look through the directions of the

Table 2. *The mean values of the atmospheric pressure gradients mb/5° latd., and surface wind speed (at 10–14 m above sea level), m/sec, at different latitudes according to the TROPEX-72 observations (July–August 1972)*

Northern latitudes (degr.)	Unfiltered data			Filtered data		
	Pressure gradient	Wind speed	Number of cases	Pressure gradient	Wind speed	Number of cases
0	1.8	7.6	120	1.6	3.9	35
2.5	2.2	7.1	260	2.1	3.6	118
5	1.2	6.3	446	1.3	3.2	209
7.5	1.3	4.7	1 092	1.0	2.7	666
10	1.3	3.3	224	0.7	1.5	136
12.5	1.2	4.6	260	1.0	2.1	118

wind and pressure gradients at different latitudes in the region of the polygons (Fig. 2). The greatest stability of wind and pressure gradient directions was characteristic for the Equator. The wind blows here almost invariably to the north-west, deviating from the median direction of 300° only within ±30°. The farther from the equator the more dispersed become the wind and pressure gradient directions, and near the Equatorial trough axis (at 10–12.5° N) they become distributed almost smoothly over the compass card, with only a slightly increasing probability around 180° and 360°. This means that the axis of the Equatorial trough changed its position fairly often here and in the period of the experiment lay north–south of the extreme north point of the ship's position on the polygon (Table 1) for about equal periods of time.

It should be pointed out that in those cases with the Equatorial trough displacement slightly to the south of the extreme north ship position, in that point some systematic mistakes in the pressure gradient computations could be easily committed. This can be seen from the example in Fig. 3, where the computed pressure gradient Γ_c is directed north, while the true gradient Γ_t should be directed south. In view of this, we tried to eliminate all such cases from the subsequent analysis, but this proved practically impossible.

The fact is that we could judge a small equatorial trough axis displacement northward or southward of the extreme north ship position and therefore also the possible mistake in the pressure gradient computation mainly by the values of the angle between the direction of the wind and pressure gradient. But if we had agreed in advance to neglect all doubtful cases when this angle exceeded some fixed value, say 90 or 120° (clock- or counterclockwise), this would have cut off the natural scattering of the values about the mean value, and so the mean value of the wind deflection angle would be *a priori* distorted. In view of this, we decided to give up analysing the calculation results pertaining to 10 and 12.5° latitudes, and carried it out for 0–7.5° N latitudes only.

The main result of our study is presented in Figs. 4 and 5. The first one gives a graph showing the relationship between the wind speed and the atmospheric pressure gradient at various latitudes. It is seen from the figure that at the Equator and at 2.5° N the air current reacts very weakly if at all to a change in pressure gradient, as computed both from smoothed and from unsmoothed data. At the 5° N and 7.5° N a great scattering field of wind speed and pressure gradients is retained, but here some direct relationship between the two variables can already be observed, especially at 7.5° N.

Fig. 5 shows the angles of wind deflection from the pressure gradient, computed for each observation period for every ship position on the polygons and then averaged over the latitudes. Root mean square deviations from the average deflection angles are also given in this figure.

Note that a defletion angle was treated as positive when the wind direction (the point where wind is blowing to) deflected clockwise from the direction of the pressure gradient.

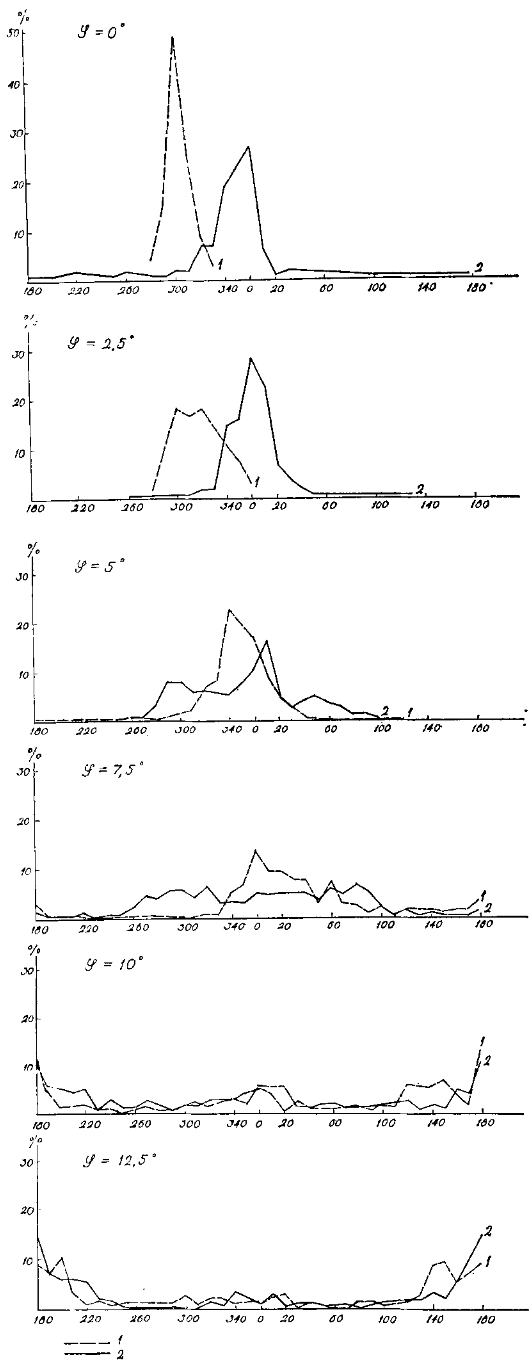

Fig. 2. The wind direction (where to) (*1*) and pressure gradient direction (*2*) distribution (raw data) at various distances from the Equator.

An opposite deflection was treated as negative, absolute values of deflection angles being less than 180°.

It can be seen that the smoothing did not change the mean angles of deflection very much. From the Equator to 2.5° N the curves for both raw and smoothed data show large negative angles of deflection. Near 6° N the deflection angles change sign, and then grow sharply up to 7.5° N. North of 7.5° N the deflection angles grow somewhat more slowly and after 10° N begin to decrease. However, this portion of the curves between 10° and 12.5° N, as we have pointed out, may prove wrong.

We believe that the presence of negative deflection angles—typical of the Southern Hemisphere—in the 0–6° N zone as well as a weak correlation between the wind speed and the pressure gradient in these latitudes, can be attributed only to the influence of the inertia of the air stream having a speed of about 7–8 m/sec here.

Note that for the whole period of work the mean directions of wind (where to) at the Equator, 2.5° N and 5° N were 304, 318 and 350° and the mean pressure gradients were directed to 337, 1 and 357° respectively. Obviously the air stream crossing the Equator cannot adjust instantly to the new and rather small (Table 2) pressure gradient and thus remains under influence of inertia. Only gradually does the inertia air stream moving northwards from the Equator lose its intensity under the influence of friction, new pressure gradients and the continually growing Coriolis force and becomes NNW, then N and finally coincides with the direction of the pressure gradient at 6° N. North of this latitude the wind begins to deflect clockwise from the pressure gradient, as should be expected in the Northern Hemisphere, under the influence of the Coriolis force, and its velocity begins to be determined primarily by the pressure gradient (Fig. 4). However, judging by the extrapolation of the curve in Fig. 5 north of 5–7° N, the deflection angle of 70 to 80°, typical of the geostrophic relation between wind and pressure fields over the sea can be observed only north of 11–13° N.

It ought to be said that the idea of an important role of the inertia in the air stream dynamics near the Equator to some extent contradicts the results of the calculation of the

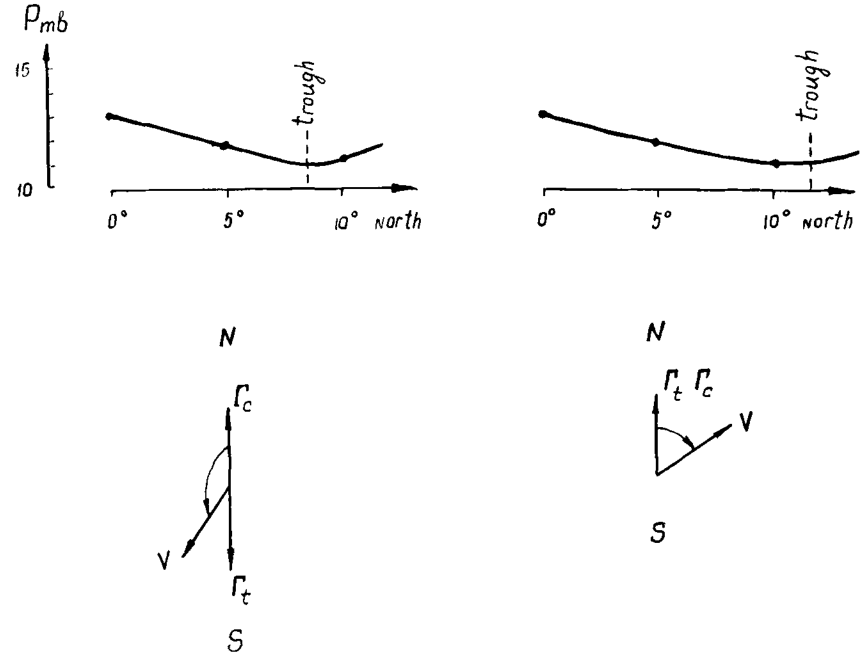

Fig. 3. Two schemes of the meridional pressure distribution and the relation between the true pressure gradient Γ_t and the computed one Γ_c. The left scheme (low) gives for ship position at 10° N an example of the discrepancy between the directions of the true meridional pressure gradient Γ_t and the computed one Γ_c, when estimated from the atmospheric pressure observations at three points: 0°, 5° N and 10° N. The right-hand scheme has no such discrepancies between the true and computed pressure gradient directions at 10° N.

values of the different terms in the equations of motion (1).

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z} - lv = -\frac{1}{\varrho}\frac{\partial p}{\partial x} + K\frac{\partial^2 u}{\partial z^2} + A\left(\frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2}\right)$$

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z} + lu = -\frac{1}{\varrho}\frac{\partial p}{\partial y} + K\frac{\partial^2 v}{\partial z^2} + A\left(\frac{\partial^2 v}{\partial x^2} + \frac{\partial^2 v}{\partial^2 y}\right) \quad (1)$$

Here l is the Coriolis parameter, K is the coefficient of the vertical turbulent exchange, and A is the coefficient of horizontal turbulent exchange. Other notations are standard.

The estimations of the values of the different terms in the equations of motion for tropic latitudes were carried out recently both on the basis of continental and island weather station data (Romanov & Romanova, 1969; Romanova & Romanov, 1970) as well as of observations from ships, including the data of the TROPEX-72 and GATE-74 expeditions (Antsipovich et al., 1975; Bruemmer et al., 1974; Falkovich & Sharonova, 1974). As we can see from Fig. 6. summarizing these results for the surface layer, the relative contribution of the inertial terms $\Delta u/\Delta t$, $\Delta v/\Delta t$ and $u(\Delta u/\Delta x)$, $v(\Delta u/\partial y)$, $u(\Delta v/\Delta x)$, $v(\Delta v/\Delta y)$ (its root mean square values are designated on Fig. 6, respectively, as $\Delta v/\Delta t$ and $v(\Delta v/\Delta s)$ does not increase towards the Equator. At all the latitudes they are roughly 10 times smaller than the pressure gradient force $(1/\varrho)$ $(\Delta p/\Delta s)$. The terms $w(\Delta u/\Delta z)$ and $w(\Delta v/\Delta z)$, too, are small in comparison with $(1/\varrho)$ $(\Delta p/\Delta s)$, as calculated by Falkovich and Sharonova, 1974, for the intertropical convergence zone. The same relationship between the above terms in the equations of motion for the equatorial region is observed in the core of the tropical troposphere

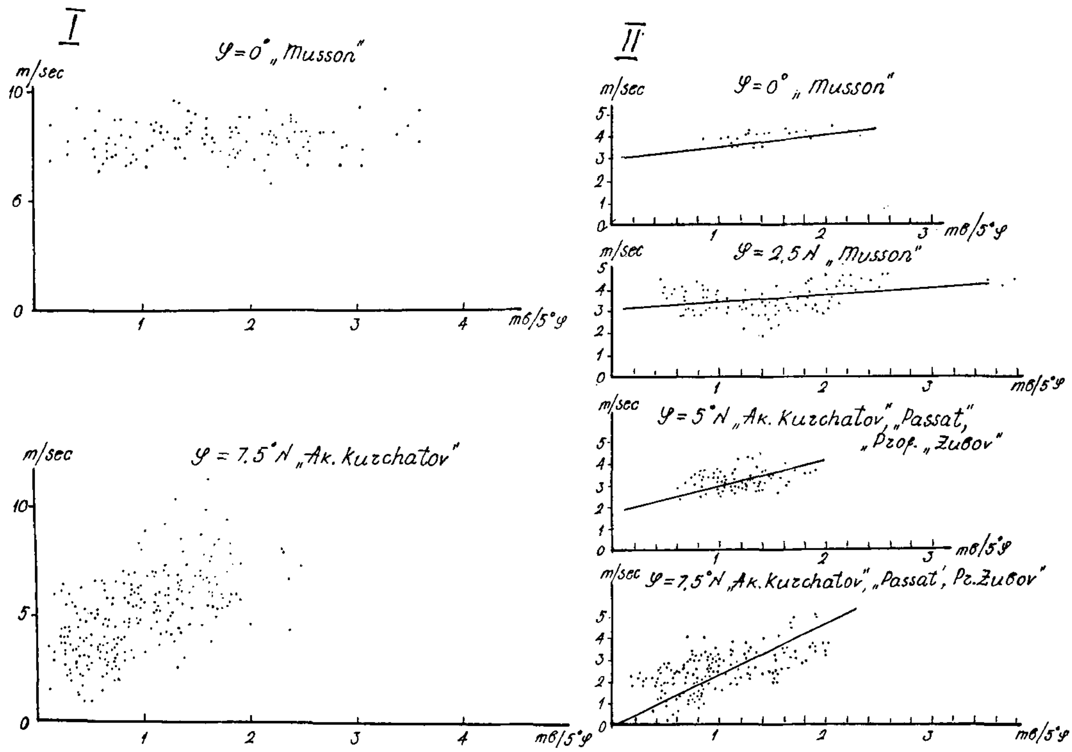

Fig. 4. The relationship between wind speeds and pressure gradients at different latitudes. (*I*) Raw data; (*II*) smoothed data.

(Antsipovich et al., 1975; Falkovich & Sharonova, 1974; Romanova, 1970) as well as near the ground (Fig. 6).

Thus the force of the pressure gradient proves to be stronger than any other force under consideration near the Equator and it appears that the surface equatorial wind must be primarily determined by the pressure gradient and the vertical turbulent exchange $K(\partial^2 u/\partial z^2)$, $K(\partial^2 v/\partial z^2)$ and have the same direction as the pressure gradient. This conclusion, as was mentioned above, is in contradiction to all our previous ones concerning the important role of inertia in the dynamics of the air stream near the Equator. We think that the correct resolution of this contradiction should include an assumption of a major contribution to the inertia from the pulsation motion rather than from the everage ones. This means that the pressure gradient near the Equator should be balanced by horizontal turbulent exchange, i.e. by the terms $A(\partial^2 u/\partial x^2 + \partial^2 u/\partial y^2)$ and $A(\partial^2 v/\partial x^2 + \partial^2 v/\partial y^2)$, which still remain to be estimated.

Some peculiarities of the wind field near the Equator can be accounted for by the inertia of the air stream. The presence of negative deflection angles north of the Equator may result in the asymmetry of the wind vorticity field relative to the axis of the Equatorial trough. North of it, where the wind deflects to the right from the pressure gradient, the wind relative vorticity will be positive (or cyclonic), but southward of the axis it may be negative.

To check this assumption we computed mean values of the relative and absolute vorticity for 0 m – 500 mb layer for the northern and southern parts of polygons separately (Table 3). These mean values were obtained for the period of intensive trough existence when its axis was located approximately along the central latitude of the polygon, close to 6° N, where deflection angles are changing sign.

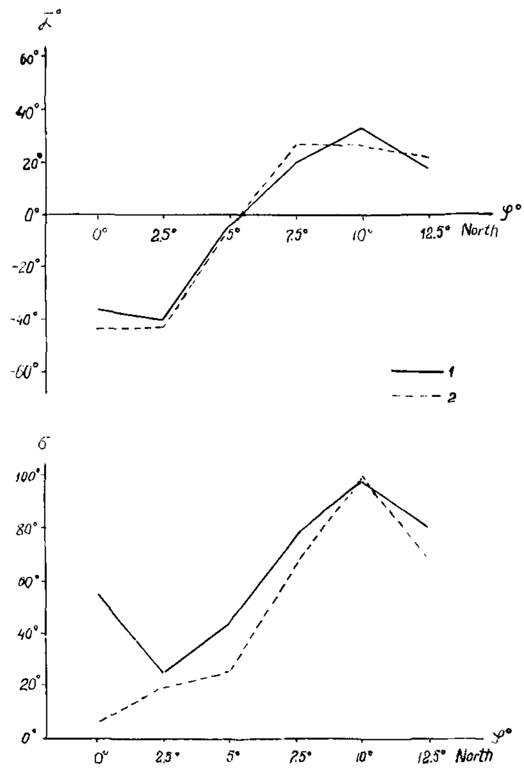

Fig. 5. Mean values ($\bar{\alpha}°$) and variances (δ) of wind deflection angles at various latitudes. The angle is positive for wind deflecting to the right of pressure gradient, and vice-versa. (*1*) Raw data; (*2*) smoothed data.

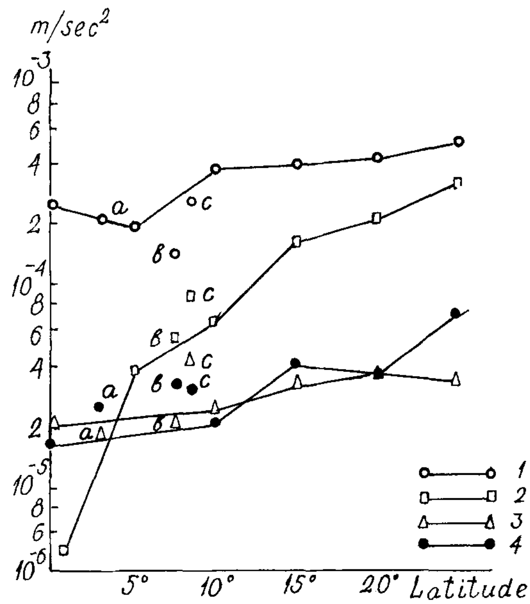

Fig. 6. Latitudinal variation of the magnitudes (mean root square values) of the different forces in the equation of motion as computed for the surface layer: (*1*) pressure gradient term ($1/\varrho\ \Delta p/\Delta s$); (*2*) Coriolis term ($lv$); (*3*) inertia terms ($\Delta v/\Delta t$); (*4*) ($v(\Delta v/\Delta s)$). The data illustrated by curves were taken from Romanov & Romanova, 1969. The data labelled with different letters were computed: (*a*) by the present authors from TROPEX-72 observations; (*b*) by Falkovich & Sharonova (1974); (*c*) by Antsipovich et al. (1975).

The table shows that for the periods selected the mean relative vorticity (ζ) does in fact change its sign and the absolute vorticity ($\zeta + l$) considerably changes its values from northern to southern parts of polygons and consequently from the northern to the southern part of the Equatorial trough.

It should be noted that because of the air stream inertia near the Equator the deflection angles might be of either sign. Should the wind in our case have at the Equator a northeasterly and not a northwesterly direction, positive deflection angles would occur anywhere north of the Equator.

The abnormal negative deflection angles north of the Equator and the abnormal positive

Table 3. *Mean relative (absolute) vorticity for 0–500 mb layer as computed for selected periods over the northern and southern parts of polygons*

Polygons...	Mesometeorological		1st synoptical	
Weather at polygons at the selected periods...	Shower	Weak rain	Shower	Weak rain
Number of upper air observations	7	17	8	32
$\zeta(\zeta + l) \times 10^{-7}$ sec^{-1}				
north part	179 (431)	74 (326)	−8 (244)	4 (256)
south part	−36 (90)	−11 (115)	−107 (16)	−10 (116)

angles south of the Equator as well as the asymmetry in wind vorticity distribution around the Equatorial trough can be observed only during a prevailing easterly wind condition. As we know, the regions with prevailing easterly winds dominate the earth's Equatorial zone. They embrace the greater part of the Pacific and Atlantic oceans and during one particular season—even the Indian Ocean. Thus the asymmetry described may be considered as a global phenomenon.

4. Conclusion

(*a*) The analysis of the TROPEX-72 observation data showed that in the narrow near-Equatorial (0–2.5° N) zone of the eastern Atlantic the air current depends very slightly on the local pressure gradient. This dependence becomes noticeable only from 5–7° N.

(*b*) It is inferred from the above conclusion (*a*) and from the estimates of the values of the different terms in the equation of motion for the Equatorial region that the inertia associated with turbulent motion can play the principal role in the dynamics of the atmosphere here.

(*c*) The inertia of the air stream near the Equator is responsible for the asymmetry of the vorticity field around the axis of the equatorial trough (cyclonic relative vorticity north of the trough axis and the anticyclonic one south of the axis in Northern Hemisphere).

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НЕКОТОРЫЕ РЕЗУЛЬТАТЫ АНАЛИЗА ПОЛЕЙ ВЕТРА И ДАВЛЕНИЯ ВБЛИЗИ ЭКВАТОРА

Метеорологические наблюдения советской экспедиции ТРОПЭКС-72 в Антлантическом океане, полученные в июле–августе 1972 г. пятью исследовательскими судами на трех полигонах, использованы для анализа соотношения между приземным ветром и градиентом давления на разных расстояниях от экватора. Установлено, что в узкой приэкваториальной 0–2,5° с. ш. зоне, где наблюдается довольно сильный 7–8 *м*/сек юго-восточный пассат, воздушный поток очень слабо зависит от локального градиента давления и на этом основании делается вывод о большой роли инерции воздушных потоков в этой зоне. Однако из подсчета порядков членов в уравнениях движения для приэкваториальных широт вытекает, что сила инерции среднего движения оказывается здесь, как и в средних широтах, на порядок меньше силы градиента давления. В связи с этим высказывается предположение о том, что основной вклад в инерцию экваториальных воздушных потоков может вноситься не средними, а пульсационными движениями, т. е. горизонтальным турбулентным обменом. Влиянием инерции воздушных потоков удается объяснить наблюдающуюся в некоторых частях океанов асимметрию величин завихренности ветрового поля относительно оси экваториальной барической ложбины,

при которой к северу от оси ложбины наблюдается циклоническая относительная завихренность, а к югу от оси — антициклональная завихренность. Зависимость скорости ветра от градиента давления становится заметной лишь с широты 5–7,5° с. ш. Близкое к геострофическому соотношение между ветром и градиентом давления, как показывает экстраполяция наших данных, имеет место лишь к северу от широты 11–13°.