SVENSKA GEOFYSISKA FÖRENINGEN

VOLUME 2, NUMBER 1 **Tellus** FEBRUARY 1950

*A QUARTERLY JOURNAL OF GEOPHYSICS*

# On the Role of the Tropics in the General Circulation of the Atmosphere

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(Manuscript received 26 September, 1949)

## *Abstract*

Contrary to older descriptions, the air currents in the middle and upper troposphere over the low latitudes are not steady but highly variable. There are trains of well-developed waves or vortices. When these disturbances combine with troughs in the polar westerlies, the resulting extended troughs often reach from pole to equator. Poleward flow of heat must take place mainly along these troughs. Energy from low latitudes is injected into the temperate zone in narrow and variable areas only.

A description of the tropical atmosphere in the layers below and above the bases of the cumuli leads to an attempt to explain the steadiness of the low-level wind field which contrasts with the high variability of the upper layers. Then the strength of the mean meridional circulation and the fields of divergence and convergence at the surface are calculated, followed by a summary of available information on the high-level circulation. The field of motion aloft varies greatly with longitude, so that inferences concerning the tropics as a whole can only be made from maps covering most longitudes. Such maps (300-mb charts) are available for a limited period down to latitude 10° N. They indicate that regions where the easterlies decrease upward alternate with regions of increasing easterlies.

## Introduction

For many years, the belief has held sway that the presence of atmospheric disturbances is peculiar to middle latitudes, whereas steady conditions prevail in the arctic and in the tropics. This conviction helped to reduce the excessive number of variables which confront the meteorologist at every turn. In the tropics, it found strong support in the apparent monotony of surface climatic conditions over wide areas, reflected mainly in small variability of temperature and wind. The opinion prevailed that conditions in the free atmosphere were as steady as those near the ground. Therefore, with the aid of few and scattered upper-air data it was held possible to deduce the correct circulation picture aloft. On any given day the upper flow should be indicative of average conditions, in contrast to middle latitudes. Solar energy accumulated in the tropics ultimately must sustain the rapid westerly flow of higher latitudes. Yet, because of the steadiness of the tropical regime, short-period fluctuations of kinetic energy outside the tropics are not traceable to variations in the

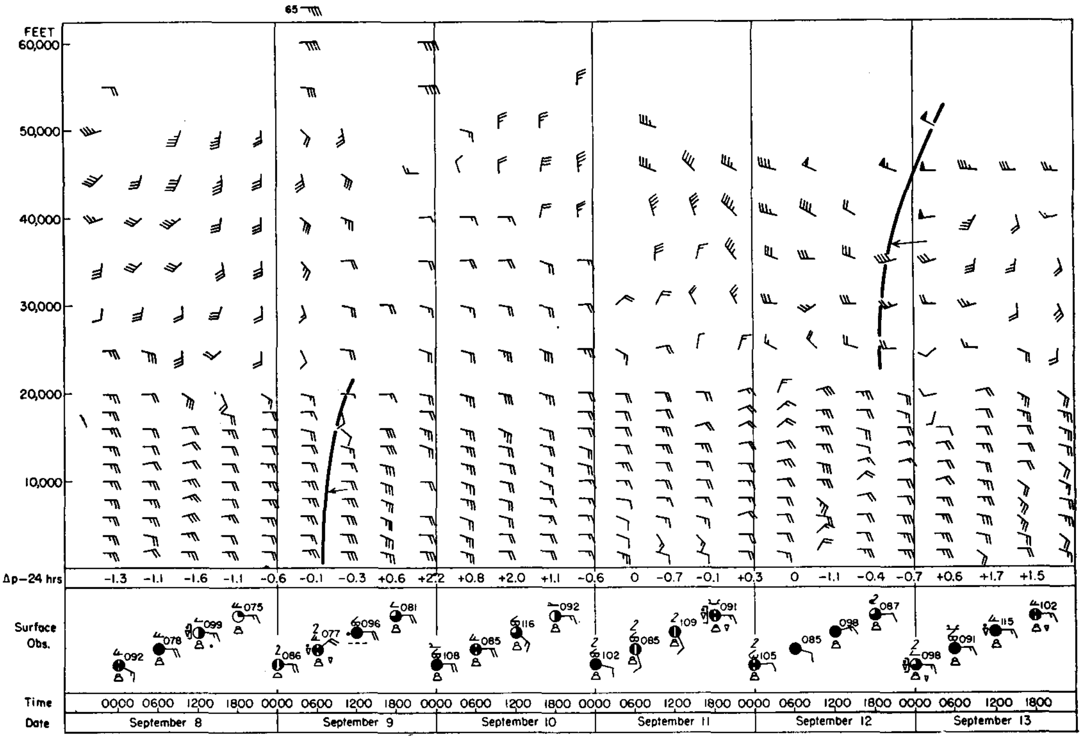


Fig. 1. Vertical time section at Eniwetok (11° N, 162° E), September 8—13, 1945 (Greenwich time). A long barb represents 10 mph, a short barb 5 mph and a heavy triangular barb 50 mph. Heavy solid lines are trough-lines. 24-hour surface pressure changes are plotted above the surface observations.

source region of the energy. Poleward transport of heat energy across the subtropical ridgeline is subject only to seasonal and long-period variations.

Gradual accumulation of aerological data, however, has revealed that space and time variations, especially of the wind flow aloft, are surprisingly large. Even the pre-war literature carries the annoyed comment that the more numerous the observations from the tropics, the more complicated tropical analysis. The large amount of high-level data collected in most parts of the tropics during the recent war finally has made it amply clear that in many low latitude regions the variability of winds aloft equals that of the temperate zone.

This is true notably in the upper troposphere. Let us glance at a time cross-section of rawin observations taken at Eniwetok in the Marshall Island group of the Pacific Ocean (fig. 1), a section typical of the sequence of events aloft over wide areas of the tropics. In the lower atmosphere, up to about 600-mb, the steadiness of the flow is undeniable. From there on upward fluctuations increase. *Extreme restlessness rules in the upper troposphere,* to subside only as we pass across the tropopause into the stratosphere. The classical notion about the tropics thus is partly right — but it gives only half of the story. In the low troposphere the course of events is very uniform, especially the important intake of heat and moisture from the oceans which is dependent in part on the surface wind. This, however, cannot hold for the poleward discharge of heat aloft. At the very least, this heat flow must vary from longitude to longitude as north and south winds alternate along the latitude circles.

A hemispheric map of the upper troposphere (fig. 2) brings this out even more clearly. For several months toward the end of the last war simultaneous upper-air observations were taken throughout the tropical and equatorial belts of the northern hemisphere: over Africa, the Americas, the Pacific Islands, and south-

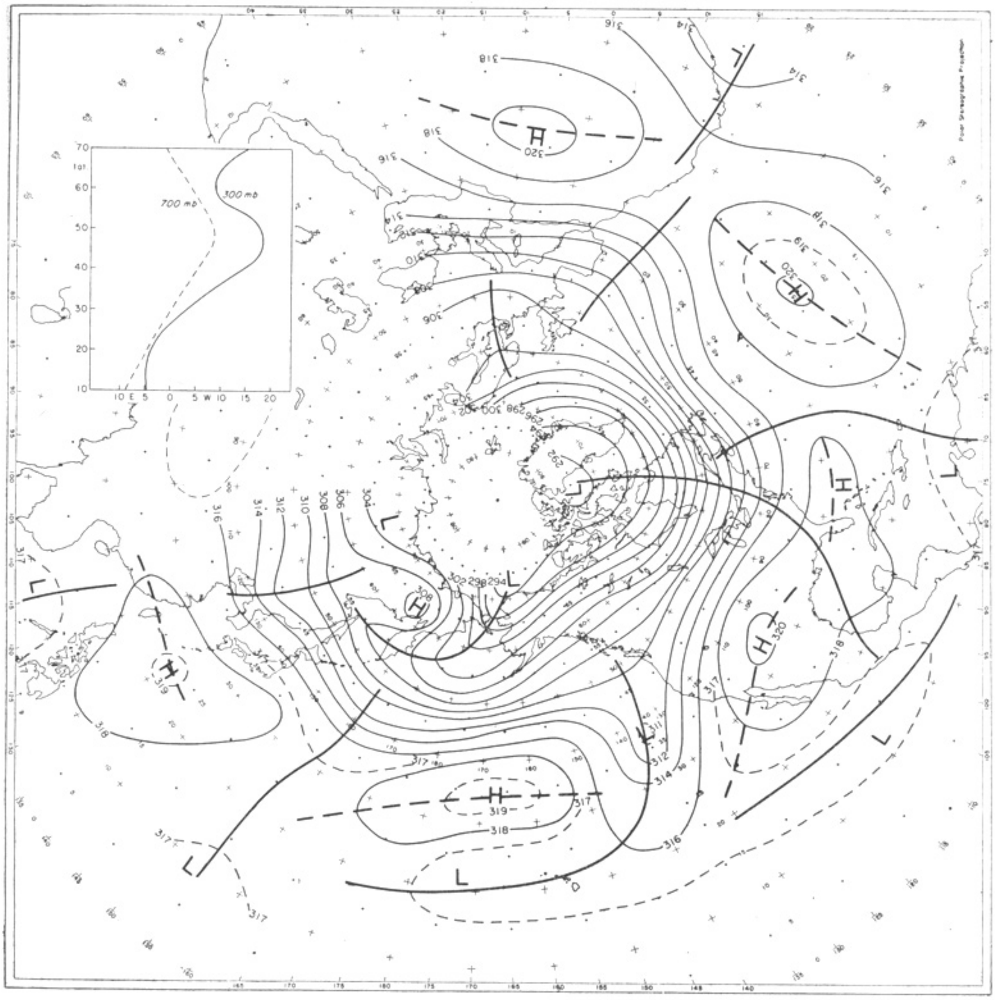


Fig. 2. Topography of the 300-mb surface (100's feet), August 26, 1945, 0600Z. Insert: meridional profile of the zonal wind (mps) at 300 and 700 mb computed between longitudes 20° E and 110° E via western hemisphere.

eastern Asia. These months are eminently suited for large-scale analysis. Under auspices of the Office of Naval Research, United States Navy, the entire period was analyzed at the University of Chicago, both at low and high levels. The chart for August 26, 1945, represents a random selection from this file. Even under the favorable circumstances prevailing in 1945 station coverage was poor in some areas, especially in the tropical Atlantic. Differential analysis therefore was a standard auxiliary analysis tool, together with time cross-sections and other techniques employed under such conditions. It is not possible to vouchsafe for all details of fig. 2. But its broadscale features are submitted without hesitation as reliable.

The most outstanding of these features are the *breakdown of the tropical atmosphere into a train of vortices* (RIEHL 1948) *and the complete interlocking of flow between high and low latitudes.* It would seem more than difficult to draw a line separating polar and tropical zones. Everywhere, troughs of great extent — called »extended troughs» by CRESSMAN (1948) — reach from pole to equator. It is only to the east of these troughlines that heat can flow poleward. Thus there is not a general seepage of energy northward, uniform in longitude. Heat is injected into the polar zone in few and narrow strips of longitude. In part, therefore, changes of flow configuration and intensity in higher latitudes must depend on the availability of low latitude disturbances to form extended troughs. We can substantiate this point further by calling on an entirely different source of evidence.

**Composition of tropical rainfall**

The rate of poleward flow of heat energy along extended troughs depends partly on the local energy supply in a trough, and the energy that can be drawn advectively toward it. Since the free atmosphere is a cold source with respect to radiation, at least up to 200-mb, condensation heat, latent or converted, is the only heat energy form available for transport. If we should find that the lower atmosphere transmits such energy to the upper levels with preference along troughlines in low latitudes, it would follow that extended troughs—whose existence is transitory—are far better heat conductors than would be the case if advection from large and distant areas were required.

As the bulk of the low latitude precipitation falls in the form of showers, convection is the prime mechanism for distributing condensation heat vertically. It has been the custom to portray convection as occurring predominantly with a random spatial distribution, limited only by climatic boundaries. This picture follows from the conception of a steady tropical circulation. Slight rainfall variability should accompany the minute fluctuations of surface temperature and wind. As is well known, this does not hold true, even from the seasonal viewpoint. In any given area, excess rainfall may flood the fields in one year and drought may strike in the same season of the next year. If we determine in what manner the individual days contribute to the rainfall total, the discrepancy with the classical description becomes even more striking, even though we hear that in many locations showers begin every day at a certain hour with clocklike regularity, and even though the Hawaiian tongue contains no word for "weather".

Along with many others who have ventured to dispute the classical viewpoint, this writer (1945) has maintained that rainfall of the tropics is very variable from day to day. At any location clear and rainy spells alternate, and this alternation occurs in definite relation to changes in the upper windfield. Bad weather concentrates in narrow zones near troughlines. Here we observe high relative humidity to great heights, therefore the air mass usually labelled as "maritime tropical." Elsewhere, the atmosphere is divided into a lower moist and upper dry layer, with a boundary between 5,000 and 10,000 feet depending on location and season. This boundary exists even where the trade wind inversion or layer of thermal stability are absent. *Thus, there are good qualitative indications that the upper troposphere receives its latent heat supply in large measure in disturbed zones.* Moreover, CRESSMAN (1948) has demonstrated that the amplitude of high and low latitude disturbances and the intensity of the associated bad weather increases whenever these disturbances join to form extended troughs.

These statements are susceptible to quantitative study. The existence of skewness in the rainfall distribution at tropical stations on a monthly or annual basis is well known. The average rainfall is the result of many periods with less than average and a few periods with heavy precipitation. This skewness becomes more pronounced and sometimes extreme if we investigate how the daily rainfall contributes to monthly or annual means. Such studies were made recently for rainfall on Puerto Rico (RIEHL and SCHACHT 1947) and for Florida[1].

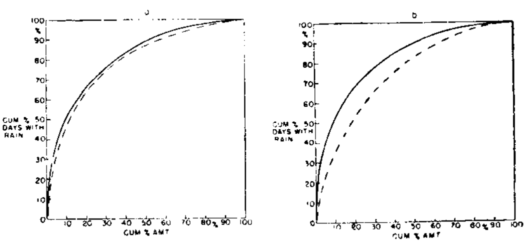


Fig. 3. Cumulative percentage frequency distribution of rainfall at San Juan, Puerto Rico (solid curves) and for average of all stations on Puerto Rico (dashed curves): a) for dry season, February—March; b) for rainy season, August—September. Ordinate: percent of days with rain; abscissa: percent of rainfall. Ten years were used to compute diagram. Hurricane precipitation is not involved.

[1] H. RIEHL: "Florida thunderstorms and rainfall." J. Meteor. **6**, 289—290, letter to editor.

Similar results were obtained for the Philippines in the course of class instruction. On the island of Oahu in the Hawaiian group over 2/3 of the annual rainfall stems from about 10 major storms which on the average last 3—4 days (RIEHL 1949). As another example, fig. 3 illustrates conditions at San Juan, Puerto Rico. This city enjoys a moderate climate with average monthly rainfall ranging from 2.5 to 7 inches in the course of the year. During both dry and rainy season 10 % of all days with rain account for 50 % of total rainfall, whereas 50 % of all days with rain produce only 10 % of the precipitation. This holds also for the rainfall of Puerto Rico as a whole, including many mountain stations.

It follows that most tropical rainfall occurs in consequence of organized, not unorganized, convection. *Within relatively few and narrow zones of disturbed weather does the tropical atmosphere actually obtain the largest share of that portion of its heat which is derived from release of latent heat.* These zones, the "secondary" disturbances of low latitudes act as tropical counterpart of the cyclones of higher latitudes in effecting conversion from latent to sensible heat. This confirms the hypothesis raised at the beginning of this section. The bulk of the heat energy transported to high latitudes along extended troughs is injected into the upper levels within these troughs.

Our attempt to demonstrate that the tropics are a factor in producing changes of the general circulation so far has yielded the following: the field of motion aloft in low latitudes is marked by great restlessness connected with the passage of vortex or wave trains; heat flows poleward in a few narrow channels, variable in longitude and time; the heat available for transport is deposited in the upper air in a few small areas. These observations also imply that it is not possible to arrive at an understanding of the general circulation by employing only steady state dynamics applied to a statistical mean circulation. This was recognized early by DEFANT (1921) for the higher latitudes. ROSSBY's studies (cf. ROSSBY 1949) place increasing emphasis on the individual members that compose the mean circulation. Outside the tropics, the flow of air is variable in space and time at all heights. It is a peculiarity of the tropics that for the most part only the high atmosphere is characterized by this variability, while the low levels are steady. Let us briefly turn to this layer with steady circulation.

## The "subcloud" layer

Perhaps the most perplexing fact about the air currents below the bases of the cumuli in the tropics is their regular monotone flow. Nowhere else is the atmosphere capable of maintaining an equilibrium state locally. The intake of heat and moisture from the tropical oceans, and the heat accumulation in the lowest part of the tropical atmosphere actually take place in steady broadscale currents, featured only by small-scale vertical turbulence. JACOBS (1942) has made comprehensive calculations of seasonal and geographic distribution of energy transfer from ocean to atmosphere. The most recent information on the subcloud layer itself comes from two Caribbean expeditions undertaken by Woods Hole Oceanographic Institution. During the first of these voyages experiments were made with smoke released from ships in order to detect any vertical and horizontal eddies that might be present (WOODCOCK and WYMAN 1946). Smoke plumes were observed to drift downstream with lateral distortion of the orientation of the

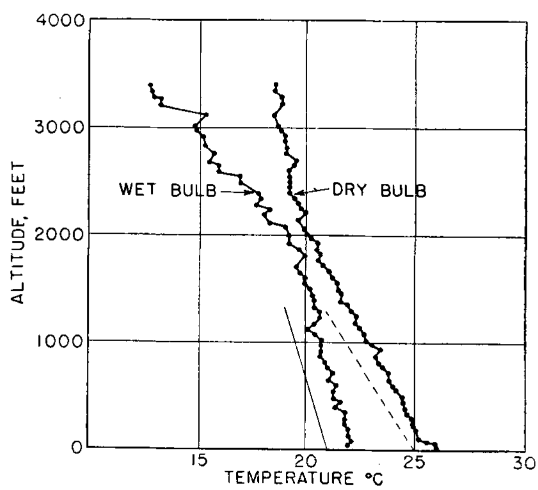


Fig. 4: Vertical structure of the subcloud layer near Puerto Rico on April 2, 1946. Airplane ascent was made about noon local time. Light dashed line is dry adiabatic line, light solid line is line of constant specific humidity. Reproduced with permission of Woods Hole Oceanographic Institution.

plumes from the mean wind direction that attained over 45°, whereas the calculated crosswind varied by 5° at most. In the vertical, smoke rose up to 150 m and successive upward bulges were about 300 m apart. This is taken as evidence that the plumes are floating in a convection cell pattern that with simplification can be patterned on the Bénard cell.

The second expedition, led by J. Wyman and A. H. Woodcock, took ship and aircraft measurements of temperature, humidity, turbulence, and vertical motion from the vicinity of the sea surface up to the trade wind inversion near Puerto Rico during April 1946. A study of the trajectories of the air sampled showed it had reached the Antilles on paths mainly from the north-central Atlantic (BUNKER, HAURWITZ, MALKUS, STOMMEL 1949). Certain differences compared to the data of the "Meteor" expedition therefore should be expected. In particular, the large superadiabatic lapse rates of the eastern Atlantic should have disappeared. This was indeed observed. The subcloud layer near Puerto Rico turned out to be almost completely mixed (fig. 4) and can be regarded as such for all large-scale atmospheric problems. In the upper half of the layer the potential temperature increases very slowly with height (about 0.2°C/100 m). Conversely the mixing ratio decreases. As these gradients have inverse effects on the stability, the virtual potential temperature also was calculated. This quantity behaves as the dry potential temperature. Thus the upper part of the layer has very slight stability with respect to dry-adiabatic vertical displacements, except where the small-scale horizontal temperature gradients treated by PRIESTLY and SWINBANK (1947) are of importance. In the lower portion, vertical gradients could not be detected. The temperature at ship's deck level, however, always was lower (about 0.5°C) than the surface water temperature.

The question arises whether the observed state of the layer which is essentially steady with respect to temperature and moisture is due to convection or diffusion. As already seen, the Panama smoke experiments favor convection as does the temperature difference between ocean and deck level. Yet the upward increase of virtual potential temperature in the upper part of the layer speaks against this solution. Moreover, turbulence records taken

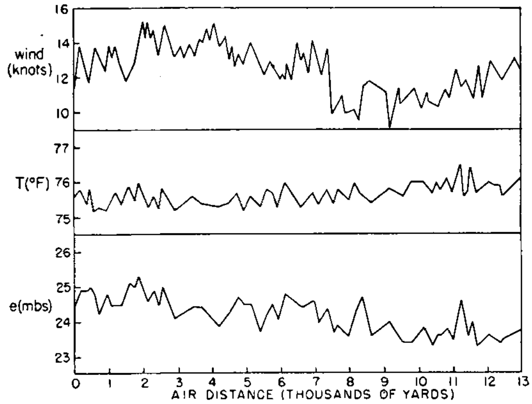


Fig. 5. Fluctuations of surface wind, temperature and vapor pressure on April 23, 1946, as obtained on ship stationed near Puerto Rico. Abscissa gives time scale converted to horizontal distance to indicate eddy sizes. The variations of vapor pressure are caused by fluctuations of water vapor density as the temperature effect is very small. Reproduced with permission of Woods Hole Oceanographic Institution.

in horizontal traverses showed only small turbulence gradients, and not even underneath the clouds themselves was there an increase of turbulence. These latter observations support the alternative, that diffusion is the governing mechanism. This is also suggested by the fact that throughout most of the subcloud layer the small-scale fluctuations of temperature and humidity mainly are out-of-phase (LANGWELL 1948) whereas at deck level they are in phase (fig. 5). In a convective atmosphere positive correlation should prevail generally.

Fig. 5, however, also indicates presence of larger eddies. Superposition of a large eddy on more minute fluctuations is easily discernible. The possibility, if not the likelihood remains that vertical motion gradients exist even in absence of turbulence gradients. If we postulate the presence of patterns of horizontal convergence and divergence that have the dimension of convection cells, it is still possible to seek the "roots" of the cumuli in the subcloud layer as a whole.

The Woods Hole observations point to a conceivable explanation for the steadiness of the windfield in the subcloud layer. Rossby has emphasized[1] that the development of large lateral turbulence eddies depends on the ther-

[1] Verbal discussions with the writer. The arguments based on the relation between vertical and horizontal eddies are included with permission of Professor Rossby.

mal stratification of the atmosphere. As the field of motion always seeks to attain gradient wind equilibrium, pressure gradients that can sustain the windfield must be able to develop aloft. On account of hydrostatic considerations this is possible only in the presence of horizontal temperature gradients that can be intensified or decreased. When lapse rates tend toward the adiabatic, vertical circulations predominate and preclude the formation of horizontal temperature gradients in the air over the oceans that are in excess of the temperature gradient of the sea surface. The lateral eddies then must recede in importance. This is precisely the situation that exists in the trade wind belt. A hypothesis can be formulated stating that because of lack of thermal stability over millions of square miles, extreme smallness of oceanic temperature gradients, and effectiveness of vertical turbulence, horizontal circulations will be held to a minimum. Flow in the low levels should be quasi-uniform, especially under trade wind inversions, since they shield the low troposphere from any disturbing influence of the high levels.

**The cloud layer**

The role of the tropical cumuli in transporting heat upward has held a place of prominence in papers on the general circulation for many years. It forms a cornerstone of most theories on that subject. Most reasoning is patterned to conform to the principles of the "parcel method" as prime mechanism of convection. The air that has travelled equatorward from the polar zones near the surface is thought to rise in random convective motion in the tropics. These buoyant parcels carry heat even to the high troposphere to compensate for radiation losses.

In recent years, dissatisfaction with the parcel method has become widespread. In particular, the observed temperatures inside cumuli are far colder than predicted by this approach. Modification of the theory of convection, begun with introduction of the "slice method" by J. Bjerknes (1938), has been carried forward recently in a brilliant paper by Stommel (1947). He explains the observed temperature and humidity lapse rates in cumuli with the assumption that a buoyant jet entrains air from all sides during its ascent.

Apart from the dynamics of convection, it is relevant to ask whether random convective heat transfer, as utilized in the older general circulation theories, is an effective mechanism under any circumstances. Consider a mass of air $M_1$, contained between the top of the subcloud layer and, say, the 300-mb level. This air is to be maintained at an equilibrium mean temperature $T$. If $M_2$, with mean temperature $T_2$, is the mass of air added from below in the course of one day to offset radiational heat losses, the total mass ($M$) aloft after 24 hours must be the sum of $M_1$ and $M_2$. Mass continuity is provided following classical theory by low-level inflow from higher latitudes and poleward outflow aloft. From the law of conservation of energy it follows that

$$MT = M_1\,(T - \Delta T) + M_2 T_2,$$

where $\Delta T$ is the heat loss in 24 hours. The ratio

$$M_2/M_1 = \Delta T/(T_2 - T),$$

where $T_2 - T$ measures the latent heat of condensation realized in the course of one day. Let $T_2 - T = 5°$ C, a large value seldom realized even in parcel method computations over the pressure interval considered, and let $\Delta T = 1°$ C, a moderate amount to judge by what is known on radiational cooling in the troposphere. Then $M_2/M_1 = 0.2$. As the bases of tropical cumuli lie near the 950-mb level, $M_1 = 650$ g per unit area and $M_2 = 130$ g, corresponding to a layer of vertical depth of 130 mb, or roughly twice the total depth of the subcloud layer.

Since vertical mass movements of the magnitude just computed must take place over the entire tropics, it is easily anticipated that the strength of the average meridional circulation will not be capable of maintaining mass continuity. In the following section the mean meridional circulation will be computed for the surface. If the result is valid for the whole subcloud layer, the ratio of meridional flow observed to that necessary to maintain heat balance according to the foregoing is $^1/_{10}$ to $^1/_{20}$. *It follows that the average equatorward flow cannot be a factor of any account in providing a mechanism for heat transport from lower to upper troposphere.* We must fall back on turbulent eddy motion.

There are two principal choices for the dimension of these eddies. They may have the size of convection cells where the vertical thermal stratification gives rise to the motion. Or they may be "secondary" disturbances where the field of vertical motion is controlled by the horizontal. The answer has already been given. Yet, as the tropical troposphere in its entirety is unstable with respect to moist adiabatic ascent — apart from trade wind inversions — we might have expected a random distribution of cumulonimbi all over the tropics, surrounded by clear areas of descending air. If this were observed generally, uniform motion might prevail at high levels as it does near the ground. Actually the closely spaced trade cumuli have their tops between 6,000 and 10,000 feet. Above this level clouds occur rather strictly in "organized convection zones", although with "Feinstruktur". *It is a remarkable coincidence that the steadiness of the trade begins to decrease upward rapidly just above the tops of the trade cumuli.*

STOMMEL (1947) to date has offered the only acceptable explanation for the constantly observed restraint upon the growth of cumuli in spite of existence of positive areas on thermodynamic charts that often are huge. AUSTIN (1948) also has pointed out that because of entrainment large density gradients between cumuli and their surroundings should not occur. An explanation for the calculated entrainment rates has not yet been offered. Yet they are evidently of sufficient magnitude to prevent growth of random convective cells to great heights all over the tropics. The average height of the tops of trade cumuli outside of areas with strong trade wind inversion must be indicative of the rate of conversion of heat and kinetic into potential energy by entrainment. Because of this conversion upward heat transfer in vertical convection cells does not carry to the upper troposphere. This compels the development of an entirely different class of disturbances —large-scale horizontal eddies in the form of vortex trains—which funnel heat aloft in zones of organized convection. The vertical stability necessary to permit lateral concentration of isotherms aloft to support the windfield is produced by the small-scale lateral turbulence in the low levels that finds its expression in the form of entrainment.

**The average circulation at the surface**

We shall now carry out the calculation of the strength of the average meridional circulation at the surface. For this purpose, the extensive summaries of resultant winds con tained in the Atlas of Climatological Charts of the oceans are entirely suitable. The region selected extends from 25° N to 25° S. About 80 % of this area is covered by water (fig. 6). Since the mean circulation over the continents, especially Africa and Australia, resembles that over the oceans (cf. BRUNT 1939, pp. 14—15), it is reasonable to conclude that the following diagrams are valid for the whole equatorial belt.

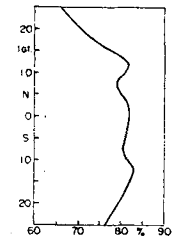


Fig. 6. Percent of circumference of globe covered by ocean between latitudes 25° N and 25° S.

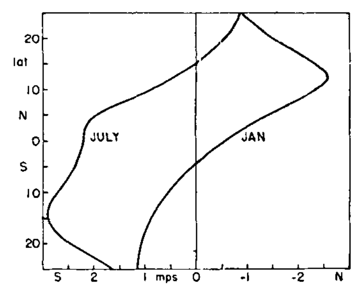


Fig. 7. Average meridional wind speed (mps) at the surface between latitudes 25° N and 25° S for the area given by Fig. 6.

The strength of the meridional flow (fig. 7) which indicates the net non-geostrophic mass transport across the latitude circles, is 1—2 mps or 1—2° latitude per day. Air crossing the 25th parallels N and S will require 2—3 weeks to reach the equator. It is relevant to ask whether the curves of fig. 7 represent small residual differences between alternating regions of northward and southward motion around the circumference of the globe on any latitude circle. Figs. 8 a and 8 b show that for the most part this is not the case, especially in July. In that month, winds with south component cover the entire oceanic area almost to 10° N, and the speed of the south component exceeds that of the north compo-

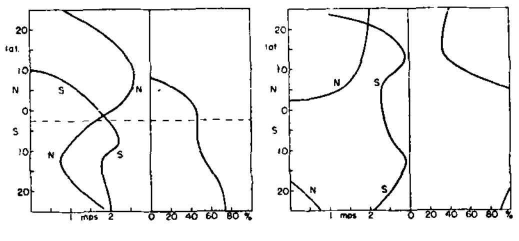

Fig. 8 a. North and south components of surface wind (mps) averaged on each latitude circle for all areas with north and south component separately: January. Right hand portion of diagram indicates percentage of circumference on each latitude circle that has south wind.

Fig. 8 b. Same as Fig. 8 a, for July.

nent. Flow of mass into the northern hemisphere amounts to about 1 % of the total mass of a hemisphere in 5 days, if the inflow extends through a layer 500 m deep.

The boundary between net transport from north and south lies near 2° S in January and shifts to 15° N in July. This displacement furnishes perhaps the fairest measure of the seasonal shift of the equatorial axis of the general circulation. It amounts to about 20° latitude whereas the sun oscillates over twice this latitude range. This fact has been taken by ROSSBY (1949) to suggest that the low latitude circulation pattern cannot be altogether thermally determined. Dynamical modifications must restrict the seasonal circulation changes to a smaller latitude range than the solar movement.

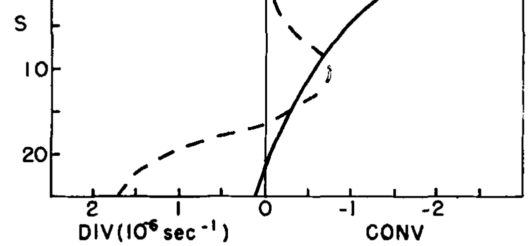


Fig. 9. Average velocity convergence and divergence from 25° N to 25° S for January (solid line) and July (dashed line). This diagram was computed by differentiation of Fig. 7 with respect to latitude.

Fig. 7 bears out the contention, made earlier, that the average meridional flow is an insignificant factor for the vertical heat transfer. Nevertheless, the computed meridional velocity gradients are quite appreciable. This can be seen if the curves of fig. 7 are differentiated with respect to latitude to obtain the field of divergence and vertical motion (fig. 9). The magnitude of the divergence is $10^{-6}$ $sec^{-1}$. Vertical displacements of 300—500 m per day would result at 3 km if the divergence was uniform between the ground and this level. These values are excessive. It is reasonable to suppose that the meridional wind component weakens rapidly in the friction layer, a contention supported by windroses at many stations. If the divergence decreases linearly with height and becomes zero at 900 mb, the vertical motion through the 900-mb surface has the order of 100 m/day, a reasonable value.

The principal region of convergence and upward motion lies in the northern hemisphere in both seasons. It experiences only a slight latitudinal shift and a surprisingly small change of intensity in the course of the year. A pronounced secondary convergence zone is evident in July but missing in January. It is interesting to note that surface divergence of considerable magnitude occurs in both subtropical belts during winter. Descent must be taking place in the cloud and subcloud layers underneath the trade wind inversion.

This is confirmed by a glance at the regional distribution of surface convergence and divergence (figs. 10, 11). As should be expected, a close correlation exists between the configuration of these charts and rainfall distribution over the oceans. In addition to the broad areas of divergence in the subtropics, the elongated narrow band of maximum convergence stands out that marks the equatorial convergence zone. Its intensity is almost uniform in both seasons as also noted from fig. 9. In the Indian Ocean, a pronounced secondary convergence zone extends near latitude 10° S in July. There is no trace of a split in January. In the Pacific, however, two convergence zones are evident in both seasons. Between them, the famous dry area of the equatorial

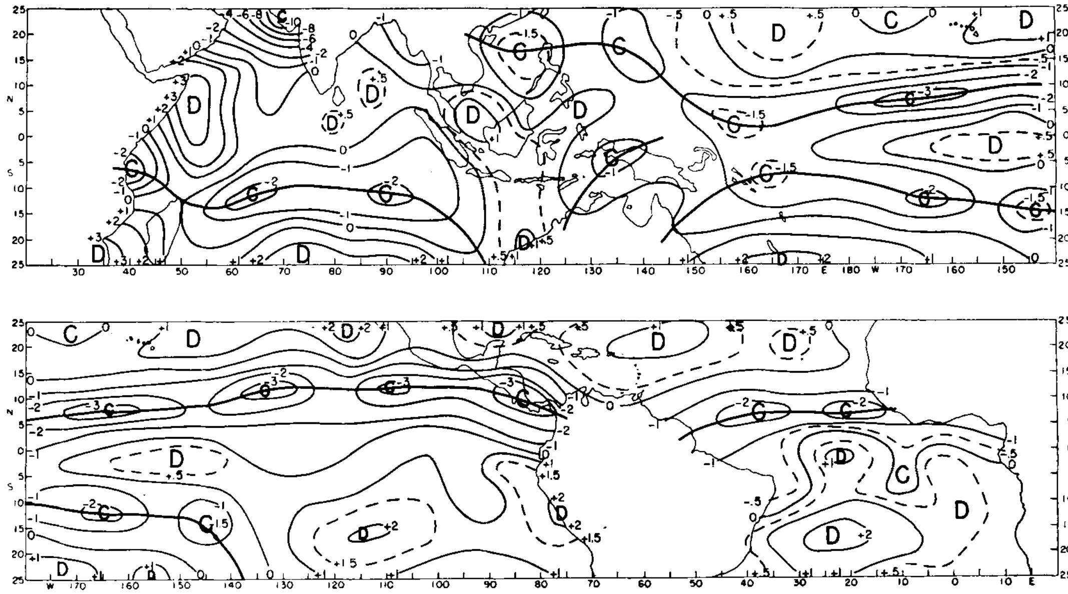

Fig. 10. Regional distribution of convergence and divergence ($10^{-6}sec^{-1}$) in January.

Pacific is very well marked. Thus the double convergence zone is a local feature of central and western Pacific during January. It extends over half of the equatorial belt in July, while the other half does not experience it as an average condition at any time.

## The circulation in the upper troposphere

Interest has centered for many years on the vertical wind distribution over the tropics. Classical theory demands that at a short distance from the equator the easterlies begin to

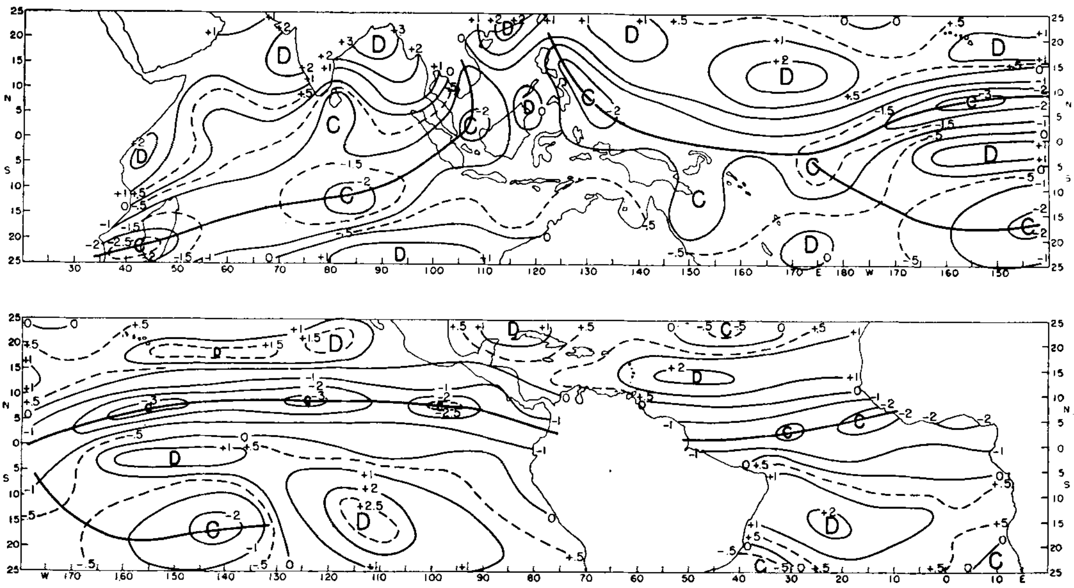

Fig. 11. Same as Figure 10, for July.

decrease with height and that a net poleward component of motion exists. In view of the preceding section, the latter requirement is undoubtedly fulfilled. It is not an object of this report to enter into the numerous theoretical arguments that are being advanced to explain the observed distribution of zonal momentum aloft. Here, we shall inquire merely concerning the state of knowledge on the actual windfield of the upper troposphere. On this point, contradictory statements have appeared of late.

Upper-wind summaries published in the course of the years have shown that in some locations the lower easterlies give way to upper westerlies during all seasons (cf. FASSIG 1933, STONE 1942). In other regions the easterlies attain their greatest strength in the upper troposphere at least during part of the year (India, Dutch East Indies, Philippines). When KUHLBRODT (1928) published his famous diagram of the vertical wind distribution over the Atlantic he stated that westerly winds replace the easterlies aloft. In the eastern Pacific Ocean, wind data at Hawaii show predominance of upper westerlies throughout the year.

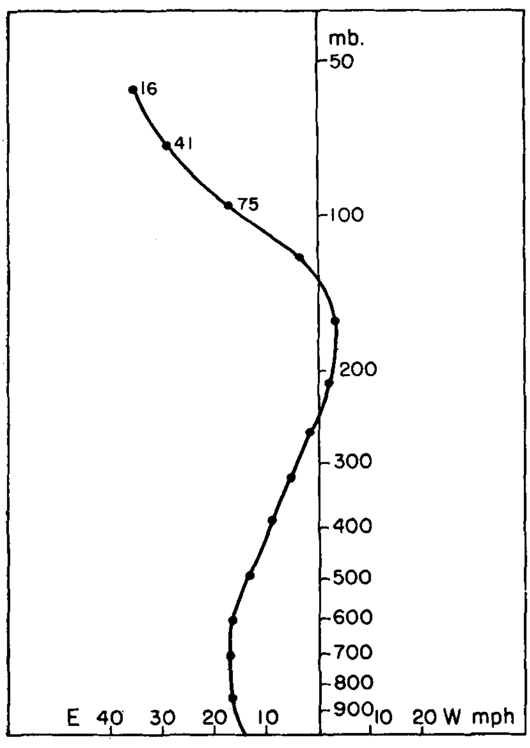


Fig. 12. Vertical variation of the basic zonal flow in the Pacific Ocean during September 1945, as determined from rawin observations. Diagram from RIEHL (1948).

A recent summary of rawin observations over the tropical Pacific during September 1945 (RIEHL 1948) also indicated upward decrease of the easterlies (fig. 12). HUBERT (1949) found strong westerlies at 200 mb in individual situations over the equatorial Pacific (Tarawa 1° N, 174° E; Phoenix Islands 4° S, 172° W). On the other hand, computations of a mean cross-section along the meridian 80° W led HESS (1948) to state: "In low latitudes the tropospheric isotherms are practically horizontal in both winter and summer. This lack of sleonoids in the vertical plane would seem to render untenable an explanation of the maintenance of the trade winds by north-south solenoids." The section of Hess shows easterlies increasing with height in the northern hemisphere summer, especially near latitude 20° N. VUORELA (1948) also has suggested on the basis of pilot balloon and radiosonde observations taken on a return voyage from Europe to South America that there are no solenoids to drive the trades.

This mass of conflicting evidence points out clearly that it is very dangerous to draw sweeping conclusions concerning the tropics as a whole from observations taken at a few stations and over short periods only. In view of the unsteady character of the upper flow, emphasized repeatedly, data such as Vuorela's hardly lend themselves to generalizations. It is necessary to gain a view on the tropical belt as a whole in order to obtain clarification. On account of scarcity of southern hemisphere data this is not yet possible. But the series of 300-mb charts for 1945, mentioned earlier, makes it possible to obtain at least a qualitative picture to about 10° N.

One monthly mean chart—that for August 1945—has been chosen for illustration (fig. 13). In August the zonal circulation in the north and the equatorward extent of the polar westerlies attain their seasonal minimum. Any tendency toward increase of east wind with height in low latitudes should be displayed most prominently. We note on fig. 13 that even though we are dealing with a monthly mean chart, marked contour patterns are nevertheless evident. In higher latitudes, a

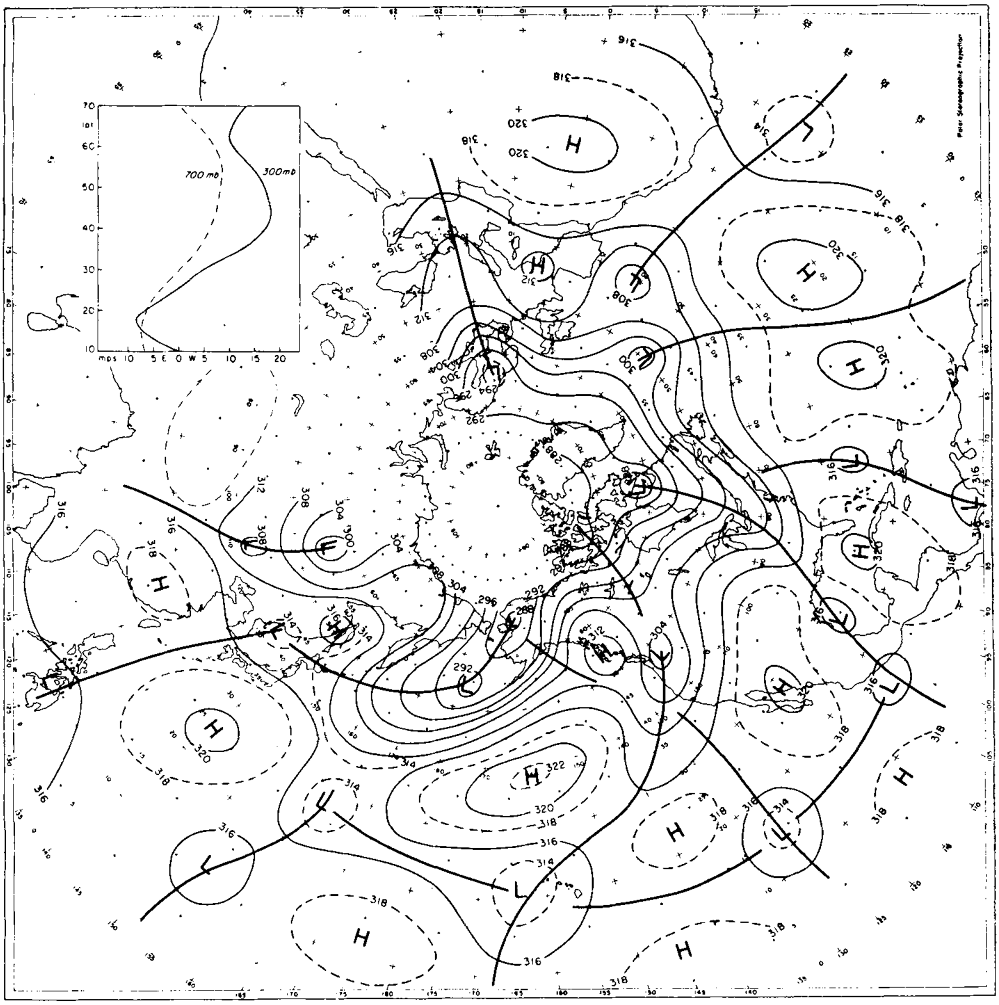


Fig. 13. Topography of the 300-mb surface (100's feet) for August 1945. This monthly chart was computed from 31 daily charts. Solid heavy lines are troughlines, dashed heavy lines mark the subtropical ridge. Insert: meridional profile of the zonal wind (mps) at 700 and 300 mb for August, 1945, computed for the area given in Fig. 2.

well-defined westerly current (jet stream) is in evidence. Pronounced troughs and ridges alternate indicating a wave number of four in the polar belt. The outstanding feature of the lower latitudes is the array of cellular subtropical highs with intermittent troughs, in good agreement with the cellular structure of the subtropics demanded by V. Bjerknes and Coll. (Physikalische Hydrodynamik, 1933, 679—687). *There is no zonal symmetry in the sense that all meridional planes show the same features.* Regions of northerly and southerly transport alternate. *A general antitrade is not present.*

All major troughs of high latitudes have counterparts in the tropics. Even on the monthly mean chart the existence of great extended troughs is the most impressive feature. It is evident, however, that the number of troughs in low latitudes exceeds that of high latitudes, in part presumably in correspondence to the much greater circumference of the earth. If in addition to the trough positions given by fig. 13 we avail ourselves of the knowledge that a very persistent upper-air trough lies near longitudes 75—80° E during summer, we obtain a mean hemispheric wave

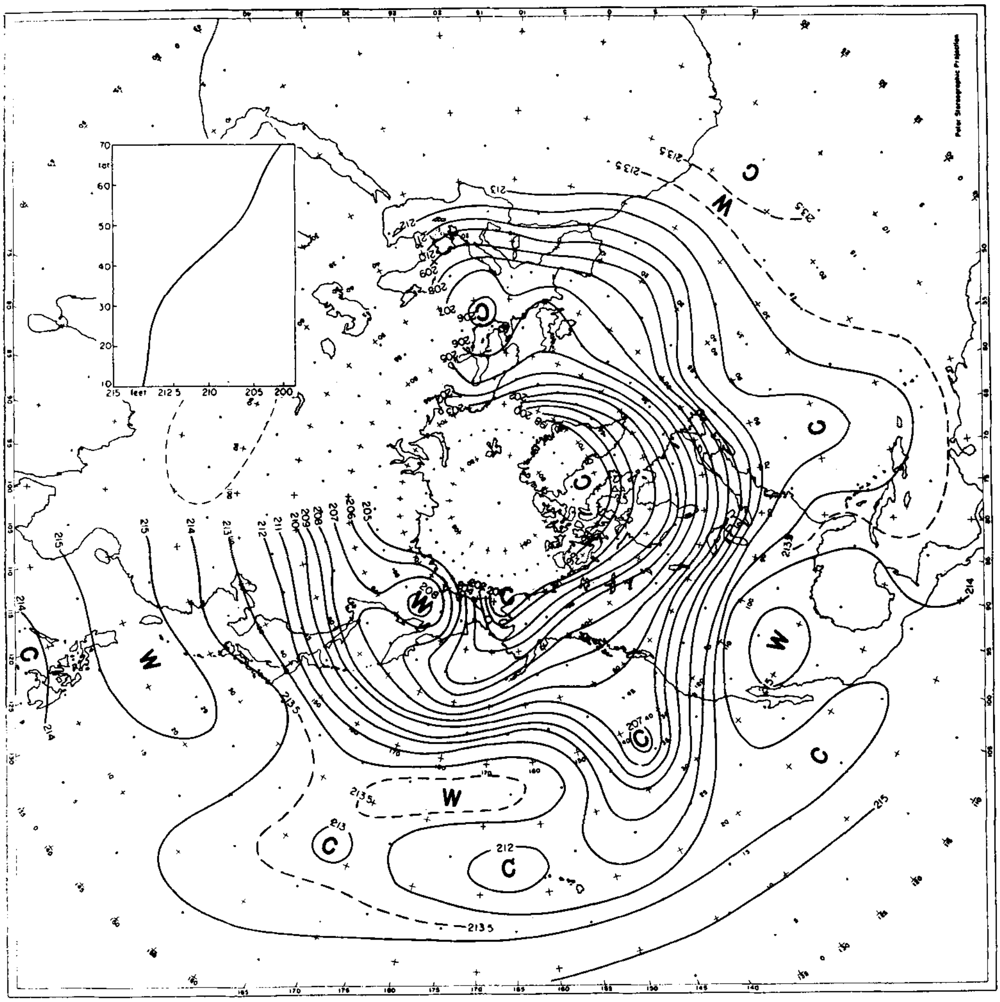

Fig. 14. Relative topography of the 300 above the 700-mb surface (100's feet) for August, 1945. "W" stands for warm and "C" for cold. Insert: meridional profile of relative topography (logarithmic scale) against latitude for the area indicated in Fig. 2.

number of seven or eight for the low latitudes. Perhaps the most remarkable asymmetry between high and low latitudes occurs in the eastern Pacific Ocean.

It is possible to compute the average zonal wind profile from fig. 13, if geostrophic conditions are assumed. Since the figure represents a monthly mean chart, this approximation presumably is valid even at low latitudes. Many observers have found that except in disturbances the deviations from balanced wind are smaller in the latitude range 10°—20° than previously assumed. Moreover, the geostrophic speeds as computed (insert of fig. 13) are in good agreement with those obtained from rawins in the tropics. Comparison of 300-mb and 700-mb shows *that the east wind decreases with height at all latitudes computed*, though it is probable from the 300-mb curve that the wind direction itself is east at that level at the equator. Even if the geostrophic assumption is only a rough approximation, the vertical change of wind must nevertheless be given correctly in the insert. Otherwise it is necessary to argue that there are systematic departures from geostrophic equilibrium that tend to produce supergeostrophic wind at 700-mb and subgeostrophic wind at 300-mb with order of magnitude of 5 mps on the monthly chart.

It is possible, however, to dispense entirely with wind calculations and merely consider the solenoid field. This can be done qualitatively from the relative topography of the 300-mb surface above the 700-mb surface (fig. 14). Around the globe there are broad areas where the temperature gradient is directed northward at all latitudes. Interspersed are regions with reversed meridional temperature gradient. Fig. 14 perhaps brings out the lack of zonal symmetry most clearly. Notably in the central Pacific and over Mexico the east wind increases with height between latitudes 20° and 30°. The same holds true over the far western Pacific south of latitude 20°. This is also confirmed by rawins, especially over the Philippines and the Palau Islands group. *Thus, there is no single meridian that will yield a representative picture of the hemisphere.* Fig. 14 is in agreement with the cross-section of HESS (1948) since along the meridian 80° W the warmest air lies near 25° N. But the contour configuration over the western Caribbean is a local feature only. A little to the east, over the Antilles, the east wind must decrease upward, in accordance with the findings of FASSIG (1933) and STONE (1942).

If the contours of fig. 14 are summarized for all latitudes, a profile results (insert of fig. 14) *that shows a northward directed temperature gradient at all latitudes.* This gradient is very small in the lowest latitude belt. Perhaps it is best to count it as zero there. As just seen, however, this does not mean that barotropy exists in any particular region. Figs. 13 and 14, together with all measured wind summaries aloft make it clear *that the average flow pattern as well as the daily chart is dominated by extended troughs and that therefore longitudinal asymmetry prevails. Regions of northward and southward directed temperature gradient alternate aloft as do regions of upper westwind and eastwind.*

Although fig. 14 demonstrates the presence of solenoids in low latitudes, it is still necessary to determine the energy source that maintains the windfield over the tropics. Sometimes it has been suggested that this energy may be derived from middle latitudes. Calculations of the heat energy released through condensation and precipitation in organized convection zones, however, show that only a negligible fraction (5 % or less) of the heat released need be converted into kinetic energy in order to account for the observed motion. This percentage is very small even in hurricanes (about 10 %). Thus there is no problem in obtaining energy to drive the currents of the tropics. The real problem is why there is so little motion, why the atmospheric heat engine is so inefficient.

**Basic currents of the tropics**

The structure of individual disturbances encountered in the variable broad-scale pattern must depend on the large-scale features themselves, especially the vertical variation of the zonal wind. The westerly current of higher latitudes almost always increases with elevation. Outside the tropics, therefore, a single type of basic current predominates. In contrast, the baroclinity of the easterlies is variable as just seen. Moreover, the tropics are subject to much larger seasonal changes of basic current structure than the higher latitudes. As a result, a greater variety of synoptic disturbances is encountered, and their occurrence has a definite seasonal course. In winter (generally dry season) the polar westerlies

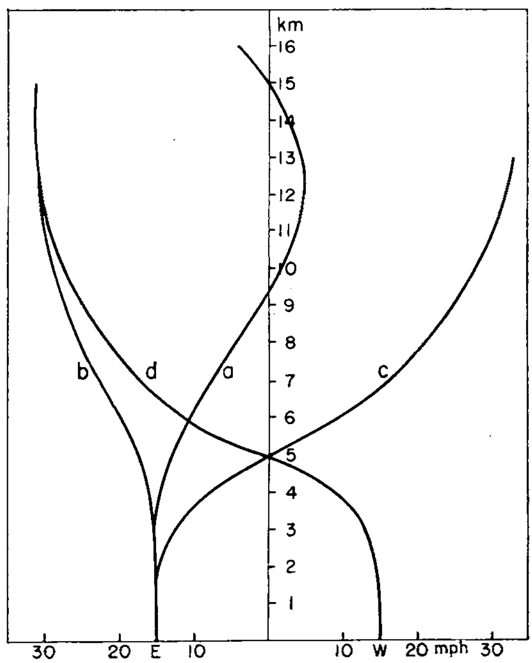


Fig. 15. Schematic representation of types of zonal currents typical of low latitudes.

aloft invade most tropical areas to at least latitude 10°. The base of these westerlies slopes equatorward with height from the latitude of the surface position of the subtropical ridgeline. Thus the depth of the trades is variable, ranging from a few hundred meters to almost the whole depth of the troposphere. Curve 'c' of fig. 15 illustrates this class of current. Curves 'a' and 'b' show the vertical variation of the trade as encountered in the rainy season (summer).

The picture is completed by introduction of the low-level "equatorial westerlies", a current that has remained least amenable to simple classification and understanding up to the present. Originally observed near the surface over India, the Far East, and West Africa, it was thought to be of southern hemisphere origin turning clockwise under the law of conservation of momentum as it crosses the equator in response to large-scale thermal differences between the hemispheres (monsoon). While this definition is not incorrect for certain areas, it is certainly incomplete. Surface westerlies in the equatorial zone not only blow toward heated continents, but also away from them to their east (RIEHL 1948). The classical type of equatorial west wind should be overlain by easterlies (curve 'd' of fig. 15). This is true in summer over India, but it is by no means observed with regularity in other regions. For such reasons it is best to avoid terminology as "southwest monsoon" that has a causal conotation. The term "equatorial westerlies", gives a definition that simply refers to the geographic fact that these westerlies are separated from the polar westerlies by an intervening wedge of easterlies. The interesting fact remains that we encounter few areas on earth where west winds do not occupy at least a portion of the troposphere.

The geographical arrangement of the four currents shown in fig. 15 follows a definite pattern. Near the poleward boundary of the tropics the trades give way to polar westerlies aloft. Low-level westerlies, overlain by easterlies, occur almost exclusively near the equatorial low pressure trough. The heat exchange between high and low latitudes of the same hemisphere largely depends on the vertical structure of the low-latitude windfield, since the latter has a definite relation to the meridional temperature gradient.[1] Poleward heat transport requires the presence of a northward directed temperature gradient, therefore a decrease of the easterlies with height. *Extended troughs form where this temperature field prevails.* As an example, comparison of fig. 13 and 14 shows that most of the principal extended troughs are situated in areas where the relative topography of the 300-mb surface decreases northward at all latitudes. In contrast, *interaction between high and low latitudes is shut off where the temperature gradient reverses.* There the easterlies increase upward to the tropopause level. *It is typical of these easterlies that they are very steady* (cf. RIEHL 1948, fig. 9). Heat energy realized from condensation presumably flows across the equator under these circumstances.

Another characteristic of the basic currents deserves mention. This is the striking opposition between lower and upper troposphere. As indicated in fig. 15, the motion that prevails near the ground differs greatly from that of the upper troposphere. A finite, generally narrow, transition zone extends between these regimes. This zone tends to be situated between 600 and 400 mb, though with many variations in individual situations. This observation points to the possibility of treating the tropical atmosphere as composed of two distinct layers in the vertical, with the 500-mb surface serving as dividing line.

As noted earlier, winds are very steady in the low levels while restlessness characterizes the high layers. There is an important exception to this statement. High-level easterlies that increase in intensity with height are very steady, as just mentioned, much more so than easterlies that weaken with height, or upper westerlies. In areas where westerlies occupy the low levels and are overlain by easterlies, eddy motion generally is strongest near the ground. The vertical gradient of wind steadiness then is reversed. It is noteworthy that a reversal of this quantity is always present. Pronounced eddies are present either above or below 500-mb. They rarely extend throughout the entire troposphere, in distinction to troughs and ridges of higher latitudes.

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[1] This argument is not based on the thermal wind equation but on empirical evidence from maps containing rawin and temperature observations. The data show that the equation is not invalid as far as latitude 10°.

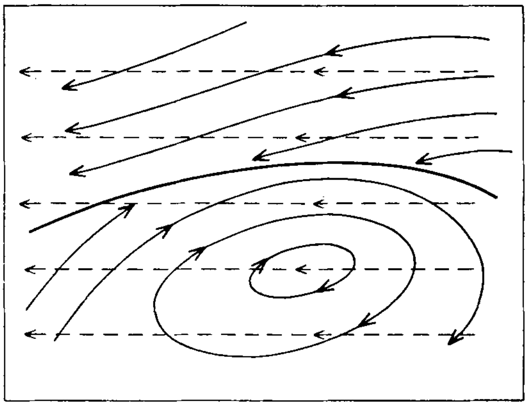

Fig. 16. Illustration showing uniform zonal current in one layer of troposphere (dashed lines) and equatorial shearline and eddy in other layer (solid lines).

Fig. 16 serves to illustrate this point. The pattern shown describes conditions frequently observed along the equatorial trough during the northern hemisphere summer. Two layers are depicted. A broad easterly current occupies one of these, while an intense shearline is located in the other. Winds are most unsteady along the shearline where cyclonic vortices develop, and to its south where motion of the clockwise eddy usually takes place. In fig. 2, conditions in the eastern and central Pacific most closely resemble this pattern. The astounding fact is that whereas fig. 16 is valid almost everywhere along the equatorial trough, it must be applied with alternating interpretation. In some regions, the shearline is located in the upper troposphere and in others at the ground (equatorial front or convergence zone). In many parts of the equatorial zone the layer of steady easterlies adjoins the surface. In others it prevails from 500-mb upward. As seen from figs. 10 and 11, however, the equatorial front zones are very narrow and occupy only a small fraction of the total tropical area.

In view of the foregoing, it is not surprising that synoptic studies report gross regional differences in equatorial weather patterns. If we recognize that a remarkable compensation exists between low and high levels, that upper and lower troposphere tend to have different and often quasi-independent flow patterns, and that the relation between these layers varies with longitude, it becomes possible to evolve a rational account of weather near the equator.

## Conclusion

This report has attempted to demonstrate that the investigator of the general circulation cannot proceed like a physicist who turns on a well-controlled flame underneath a tank with fluid and then studies the motion in the tank without further reference to the flame. The source of heat for the middle latitude westerlies has a very variable character. Transfer of heat from the oceans surface to the high troposphere—the principal layer of transmittal to the polar zone—is locally and longitudinally not continuous. This also holds for poleward heat emission. Since local and longitudinal variations of heat exchange are in part determined by the tropical circulation itself, changes of the flow pattern at the receiving end in the temperate zone cannot be understood entirely without reference to the tropics. As yet, no attempt has been made in meteorology to evaluate dynamically these poleward directed influences.

Although heat accumulation and emission from the tropics is a function of longitude, and locally also of time, the question remains whether there are fluctuations of these quantities in time if we integrate around the entire tropical belt. The wartime extension of observational material has made possible studies of this important problem. Recent research, as yet uncompleted, has shown that the question can be answered in the affirmative.[1] The total internal energy of middle and upper troposphere in low latitudes, therewith the poleward heat transport, are subject to marked aperiodic fluctuations with a period of roughly two to three weeks. This observation makes it clear that the tropics must play an important part in the production of variations in the intensity of the circulation of middle latitudes.

[1] H. Riehl, T. C. Yeh, and N. LaSeur: A study of variations of the general circulation (to be published).

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