INVESTIGATIONS ON THE TROPICAL EASTERLY JET

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ABSTRACT

During the northern summer, the Tropical Easterly Jet extends in the layer 200-100 mb in the latitude belt 5-20°N from the Philippines across southern Asia and northern Africa to the western Atlantic, i.e. over nearly half of the earth's circumference. Its persistency in position, direction and intensity is remarkable; only minor pulsations are observed with any frequency. Its kinetic energy can be derived from the equatorward displacement of angular momentum in the northern part of its entrance region (over China) together with the conversion of available potential energy into kinetic energy in the same area. Both effects are induced in time and space by the powerful high-tropospheric Tibetan anticyclone.

Based on all available actual wind observations during the period July-August 1956-62 the cross-circulations in the entrance and exit region of the TEJ are quantitatively estimated. In the exit region (above Africa) the high-tropospheric component perpendicular to the TEJ-axis is directed towards N and produces large-scale subsidence on the northern side of the jet. In the entrance region (above Southeast Asia) the high-tropospheric meridional component is directed towards S; the corresponding subsidence area can only be found south of the equator. The reversal in the direction of the cross-circulation, corresponding to the transition from acceleration to deceleration of the TEJ, is mainly produced by large-scale convective activity and vertical momentum exchange along the
west-facing meridional coasts of both peninsulae. From the thermo-
dynamic and dynamic point of view we observe therefore in the
entrance region a direct (solenoidal) cell with a
exit indirect (antisolenoidal) cell with a
work-producing Hadley cell with a
work-consuming Anti-Hadley circulation.

Each of the cells extends, with nearly equal intensity, over about
one fourth of the earth's circumference, thus reducing the intensity
of the average planetary meridional summer circulation in
tropical latitudes nearly to zero. The unique summer aridity of
the desert belt from the western Sahara to Pakistan - which
extends farther towards the equator than in all other continents -
is strongly correlated with the forced descending motion on the
northern side of the TEJ exit region. Apparently only weak
homologies of this system exist during northern summer over the
Caribbean and Central America, as well as during southern summer
over the three southern continents. In all these cases the high-
tropospheric easterlies are weak and increase with height into
the stratosphere, while the TEJ is distinctly separated from the
stratospheric summer easterlies.
1. INTRODUCTION

The occurrence of remarkably strong easterly winds above the so-called "summer monsoon" of India has been realized for only some 15-20 years. In 1948 Scherhag (32) published maps of the average height of the 96 mb and 41 mb surfaces above the northern hemisphere during July and mentioned the remarkable strength of an easterly flow above India and its neighbourhood. According to the very scanty and inhomogenous data available at the end of the second world war, he came to the conclusion that the strength of these - geostrophically deduced - easterlies increased up to the highest level in the stratosphere. For the layers between 10 and 16 kms this conclusion was confirmed by an investigation of actual winds by Venkiteshwaran (37). A revised version of all aerological data available for the 225 mb-level was given by Flohn (7), who stressed the unique height above Tibet of the isobaric levels of the upper troposphere up to 100 mbs; the average position of an isolated anticyclonic cell at 225 mb near 29°N, 98°E coincides remarkably well with more recent results (34) with a sufficient aerological network. The extremely high temperatures in the middle and upper troposphere above northern India, especially above Agra (27.2°N, 78.1°E), have been frequently discussed since A. Wagner (40), who stressed the role of the release of latent heat in the middle troposphere and the extension of convective processes up to 16 kms.
The expansion of the aerological network after 1948 and especially the gradual transition to new techniques of regular and reliable upper wind measurements encouraged studies of these strong easterlies above Southern Asia and tropical Africa. After some case studies and statistics - (2, 6, 15) - Frost (14) was the first author to draw meridional cross-sections along 40°E demonstrating the existence of a jet-like core of strong easterlies near Lat. 15°N and above 200 mbs during summer. Jenkinson (17) published a map of the 200 mb-level for July with isolated centers of easterly flow over the Bengal Sea, southern Arabia and West-Africa. Synoptic case analyses were first given by Alaka (1) over the Caribbean and especially by Koteswaram (18) over India and Africa. While above the Caribbean, i.e. south of the continent of North America, easterly jets are only rarely observed, this phenomenon is much more frequent and persistent in the Asian-African sector of the tropics.

Due to this remarkable quasi-stationarity the climatological importance of the Tropical Easterly Jet (TEJ) is much greater above easterly longitudes, and its role within the general atmospheric circulation is remarkable. In contrast to this, the high-tropospheric flow at the same season and latitude above the Pacific Ocean is characterized by a sequence of waves and vortices and sometimes even by a westerly jet (30).

This investigation will be dedicated at first to a more detailed climatological description of the TEJ in the years 1956-62, but our main concern will be a discussion of its role within the general atmospheric circulation and its relation to large-scale
synoptic and climatological patterns as studied by Koteswaram (18, 19) and discussed in a monograph by Reiter (29). It has been started during extensive, but largely unpublished studies of the Indian summer monsoon, based on daily maps and cross-sections for the years 1952 and 1956 and carried out by my collaborators. The collection of all available wind data above the Indo-Pakistan subcontinent was further extended into the whole entrance and exit region of the TEJ from Long. \(150^\circ\)E to \(20^\circ\)W, i.e. from the Marianas to the western coast of North Africa (Chapter 2). In this chapter, we also have to investigate the possible interrelations between the TEJ and the strong easterly flow in the equatorial stratosphere (10), which has been formerly discussed as the Krakatoa Easterlies. After the recent discovery of a 26-month cycle of descending rings of easterlies and westerlies in the equatorial stratosphere (28,38) we also have to check whether any biannual periodicity of this type penetrates into the region of the TEJ below the tropopause.

At all stations, actual wind observations have been investigated with first priority, due to the well-known difficulties of reliable geostrophic analysis of the high-tropospheric isobaric surfaces in tropical latitudes. Nevertheless, average meridional cross-sections of temperature at different longitudes have been prepared for comparison (Chapter 3). Following some suggestions of Koteswaram (19) a numerical investigation of the cross-circulation in the large exit region and (later) also in the entrance region of the TEJ has been made (Chapter 4). The results of this study apparently contribute to a better understanding of the
tropical branch of the atmospheric circulation (Chapter 5) and of some large-scale climatological facts which hitherto have remained in a purely descriptive stage (Chapter 6).
2. UPPER WIND CLIMATOLOGY

2.1 Average resultant winds over Southern Asia

Initially the individual upper winds at and above 12 kms or 200 mb were collected for all stations of the Indo-Pakistan subcontinent for the period 1956-62. The source of the data was the Daily Weather Record of the Meteorological Services of India and Pakistan (Tab. 1 c). With the exception of a few stations, only a small minority of the ascents reached the 100 mb-level (which coincides, in tropical areas, very nearly with the tropopause); the first results (1956-57) at this level were included in an earlier paper on the stratospheric summer easterlies (10). Average resultant winds for the layer between 12 and 20 kms are collected in Table 1, arranged according to geographical regions.

With these data, the hypothesis of a more or less regular tropical easterly jet embedded in the upper troposphere (100-200 mb) and separated from the stratospheric easterlies - as suggested by Koteswaram (18) - was investigated. Above New Delhi, where the number of data during the years 1956-7 was greatest, the resultant wind component increased substantially with height up to 21 kms (50 mb); normally at these latitudes no separate tropospheric jet could be distinguished. Furthermore the steadiness of these easterlies - as indicated by $q = v_r/\bar{v}$ ($v_r$ = mean resultant wind, $\bar{v}$ = mean scalar wind speed) - was low (46 %) at 100 mb, but increased with height (92 % at 50 mb). Only above Madras (13.1°N) and Trivandrum
(8.5°N) could a decrease of wind speed above 16 kms be found, but here only a limited number of ascents were available. The small quantity of data above 16.2 km - which is here taken as equivalent to the 100 mb-level - could not be considered as representative; a few unreliable measurements could seriously influence the results of a small sample. Slowly the number of successful ascents increased (especially after 1960), obviously to a large extent due to the improved quality of the balloons. Throughout the entire period the author was deeply impressed by the remarkable persistence of the wind direction from day to day.

From these actual wind data, several meridional cross sections were constructed. Using the mean zonal wind component $\overline{v_x}$ during July-August 1956-59 (Fig. 3), one was able to distinguish the core of an easterly jet with a resultant zonal component of more than 35 ms$^{-1}$ between 14 kms, above Madras, and 18 kms, above Bombay, i.e. right across the tropopause. In all ascents south of Lat.22°N a small decrease of the zonal component above the tropopause was found, but not at the stations of northern India. According to the much more complete data of 1961 and 1962 a core with speeds of 33-35 ms$^{-1}$ is situated above Madras and Vishakhapatnam, but now definitely below the tropopause and with a marked negative vertical wind shear above the maximum wind layer, which slopes from 16 kms at 21-23°N to 14 kms at 8°N. These results (Fig. 3) seem to be more representative and reliable than those of the 1956-59 series. They demonstrate beyond any doubt the occurrence (and persistence) of a separate tropical easterly jet (TEJ) in the high troposphere south of about 22°N, with a core near Lat.15°
and 150 mb (∼14 kms), and its separation from stratospheric easterlies.

North of Lat. 22°N the northern extension of this TEJ merges with the lower extension of the stratospheric summer easterlies, which penetrate, at these latitudes, regularly from above into the higher troposphere. Here the persistent easterlies increase substantially with height; above the Caribbean or Hawaii they normally reach 25-30 m s⁻¹ near the 10 mb-level (∼32 kms).

According to recent investigations (20), the core of the TEJ can be split into several centers; one of these centers is situated near the equator and the 200 mb-level. Unfortunately the upper wind data of Gan (Maldive Islands, 0.7°S, 73.2°E) was not available during this study, and the very few data available from Colombo (Ceylon) before 1962 had been considered unrepresentative.

According to the resultant winds of modern equipped aerological stations, the upper boundary of the equatorial westerlies or the SW-Monsoon is situated near 7 kms or about 450 mb, south of 18°N, i.e. somewhat higher than earlier estimates based on incomplete data. According to a number of time cross-sections of upper winds this boundary varies greatly from day to day, following the sequence of synoptic situations.

Above Thailand, nearly the same pattern is observed; here the core of the TEJ can be followed between 19°N (Chiangmai) to the Equator (Singapore). At all stations the resultant wind has a small but significant northerly component; the same is true at the stations of northeastern India, such as Gauhati and Calcutta. An
isotach analysis of the resultant winds at the 150 mb-level (Fig. 2) demonstrates such a northerly component at each station east of 90°E; this fact is discussed in Chapter 4.3. Two additional maps of the 100 mb and 200 mb-level were constructed but are omitted here, due to the lack of data south of Trivandrum at this stage of the investigation. The isotach axes at both levels were included in the 150 mb map: they reveal a general slope of the core from N to S in the order of 1:200. The gradual acceleration from the Pacific to India, in the entrance region of the TET, is clearly visible (cf. Chapter 4.3).
2.2 Average resultant winds over Africa

In the large exit region of the TEJ over the African continent the bulk of the upper wind data is — at least after 1957 or 1958 — much more complete and representative than over the Indo-Pakistan subcontinent. At most stations the 100 mb-level is included in routine operations, and frequently monthly resultant winds are available from such useful sources as "Monthly Climatic Data of the World". These resultant winds at the TEJ-level are surprisingly persistent from year to year (Chapter 2.3); thus individual stations with shorter observation periods could also be included.

A number of additional data was received and evaluated during the investigation; among those are Sao Tomé, Ngaoundere and Fort Lamy. These data all confirm the analysis given by Hofmeyr (16). During the first months, the data of Malakal (Sudan) were apparently biased by too low speeds; much more strongly biased when compared with Aden were the pilot balloon data from Djibouti. An unpublished series of upper winds (Habal) at Addis Ababa (1957-61) had been made available, but unfortunately the data were much limited in number above 300 mb; the vertical shear during June also suggests the occurrence of winds in the order of at least 60 knots at the TEJ-level during July and August.

The high speeds of the TEJ above Aden and Khartoum certainly supports the use of the term "Jet"; at both stations values above 100 knots are not infrequent. Above central and western Africa the gradual decrease of speed (Fig 2) is clearly visible; here we certainly cannot define this easterly current as a jet. At all
levels the isotach axes of the TEJ cross the latitude circles at small angles towards the innermost Gulf of Guinea.

In contrast to the entrance region east of 90°E, nearly all resultant winds west of 75°E have a small southerly component, at least when compared with the isotach axis of the TEJ; only south of about 5°N can a slight northerly component be computed. As above India, there exists a gradual slope of the maximum wind level towards the equator (Fig. 4). The data seem to be sufficiently reliable to permit a numerical estimate of divergence terms (cf. Chapter 4.2).

The analysis of upper winds shows that the TEJ above Africa — if we allow the use of this term also for its western exit with only weak winds of 20-40 knots — is (in the average) not a separate entity, but the continuation of the TEJ from Southern Asia. This current is maintained above the Arabian Sea, where vertical mixing processes during the monsoon season are not pronounced (cf. Chapter 4.5), but slightly shifted towards the equator.

The total length of the TEJ-system may be estimated by measuring the length of the 30 knot-isotach: we find that it reaches over some 145° of longitude or more than 16 000 kms. Even the 50 knot-isotach extends over 80°Long. or 9 000 kms. Since we cannot discard the entrance and exit regions, we must conclude that the whole TEJ-region extends over about one half of the earth's circumference, with a marked regularity in season, height, intensity and direction. These facts deserve much attention from the climatological, the synoptical as well as from the dynamical point of view.
2.3 Interannual variability

After the discovery of a nearly biannual period in the equatorial stratosphere — alternating circumfluent rings of persistent easterly and westerly winds penetrate downwards with a period of about 26 months (28,38) — it seems necessary to check the relationship between this far-reaching phenomenon and the seasonal occurrence of the TEJ. Such an investigation necessitates, at first, a comparison of the intensity, direction and variability of the TEJ in neighbouring years. In Tab. 2 comparative values of the direction (α) and the velocity (v_r) of the resultant wind, together with its constancy q = v_r/\bar{v} (\bar{v} = mean scalar wind speed) and the number of available measurements has been collected. From this collection we may conclude, that the TEJ is — at least in its core — remarkably constant from year to year. Only at equatorial stations — like Singapore (or Sao Tomé) — are the winds at the 100 mb-level subject to great variations from year to year. Obviously the downward motion of the equatorial wind rings in the stratosphere is strongly damped in the vicinity of the tropopause, where only weak interference between the strictly seasonal phenomenon of the TEJ in the upper troposphere and the non-seasonal motions of the stratospheric systems can occur. Since both periods are apparently independent, and since the average length of the stratospheric cycle (at a given height) is somewhat longer than two years, we cannot expect a clear coincidence between both features.

The remarkable interannual constancy of the TEJ as a seasonal feature leads to the conclusion that the physical conditions re-
sponsible for its development are also constant and only weakly affected by the large-scale anomalies of the general atmospheric circulation. These physical conditions are to be considered as "climatic" and nearly independent of the perpetual fluctuations of the weather; they will be discussed in chapter 4.5.
3. MERIDIONAL TEMPERATURE CROSS SECTIONS

3.1 Critical remarks on the temperature data

At present, every large-scale comparison of the temperature distribution in the upper troposphere and stratosphere is seriously hampered by systematic instrumental differences and the lack of a generally accepted standard radiosonde. This is especially true in tropical regions, where the horizontal gradients of temperature are frequently of the same order as the instrumental differences. The application of instrumental corrections as derived from the WMO-organized radiosonde comparisons at Payerne is hardly possible, due to the frequent changes in instrumentation, evaluation techniques and correction regulations in individual meteorological services. In some countries - such as the Chinese Peoples Republic including Tibet - no details nor comparisons of the radiosondes in use are known; other countries such as India use more than one type of radiosonde in their network. Therefore in these regions the use of radiosonde data of different type for the evaluation of geostrophic winds is subject to large (and frequently prohibitory) errors. From many experiences of this sort the author has been forced to use in this investigation mainly actual wind data which, however, are not unbiased by other deficiencies.

In order to obtain an idea of the - real or instrumentally caused - time fluctuations of temperatures aloft, we give a few comparisons
from New Delhi since 1949, using for comparison also the excellent night-time registering balloon ascents made before 1940 at Agra, which is less than 200 km from New Delhi.

Average temperatures for July-August

<table>
<thead>
<tr>
<th></th>
<th>850</th>
<th>700</th>
<th>500</th>
<th>300</th>
<th>200</th>
<th>100 mbs</th>
</tr>
</thead>
<tbody>
<tr>
<td>Agra 1927-40 Reg. balloons</td>
<td>22.4</td>
<td>13.6</td>
<td>+0.2</td>
<td>-24.8</td>
<td>-48.1</td>
<td>-78.4</td>
</tr>
<tr>
<td>New Delhi 1949-51 Radiosondes</td>
<td>23.6</td>
<td>12.4</td>
<td>-2.4</td>
<td>-27.2</td>
<td>-49.6</td>
<td></td>
</tr>
<tr>
<td>&quot; 1952-56 &quot; &quot;</td>
<td>23.1</td>
<td>12.8</td>
<td>-1.4</td>
<td>-23.5</td>
<td>-43.4</td>
<td></td>
</tr>
<tr>
<td>&quot; 1957-62 &quot; &quot;</td>
<td>24.5</td>
<td>14.3</td>
<td>-0.2</td>
<td>-23.4</td>
<td>-44.0</td>
<td>-73.6</td>
</tr>
</tbody>
</table>

The remarkable increase of the temperatures in the upper troposphere - which already has been interpreted as gradual warming - should be discussed with great care. Some other examples may support this caution, especially a zonal cross-section along 25°N (Tab. 4 a). Here the temperatures at and above 500 mb at the Indian stations are 2-5° warmer than both neighbouring areas with other instrument types (Karachi with a modified US-sonde). If this difference would be real as a whole, we had to expect strong southerly geostrophic components between Karachi and Jodhpur as well as strong northerly components between Guwahati and Tengchung. In contrast to this, zonal wind components are in any case much larger than meridional components. This indicates that the real zonal temperature gradients ought to be smaller than meridional gradients. A comparison between the high-tropospheric temperatures above eastern Africa (Tab. 3 c) - using
radiosondes from the U.K. and Finland — and those above central and western Africa (with French equipment, Tab. 3 d) indicates similar differences above 300 mb ¹).

Using internally consistent ascents we can easily find that the real time fluctuation of summer temperatures in tropical latitudes are small, nearly as small as at the equator itself. As an example the following values for Bangkok (13.7°N) are given:

¹) According to the Payerne comparison of 1956 (cf. P. Beelitz, Meteor. Abhandl. Fr. Univ. Berlin VII, 4, 1958) some of the systematic temperature differences between radiosonde types reach 2–5°C and lead therefore to considerable inconsistencies. The following table contains the temperature differences averaged for 300, 200, 100 and 50 mb between the USA-radiosonde (as reference) and other types ( x – US ), separately for daytime and nighttime ascents:

<table>
<thead>
<tr>
<th>Radiosonde</th>
<th>day</th>
<th>night</th>
</tr>
</thead>
<tbody>
<tr>
<td>United Kingdom (Kew II)</td>
<td>-0.6</td>
<td>+0.3°C</td>
</tr>
<tr>
<td>Finland (Väisälä)</td>
<td>+0.4</td>
<td>+0.4</td>
</tr>
<tr>
<td>Japan</td>
<td>+0.7</td>
<td>+0.7</td>
</tr>
<tr>
<td>France (Metox)</td>
<td>+2.6</td>
<td>+0.6</td>
</tr>
<tr>
<td>India (Chronometer)</td>
<td>+2.3</td>
<td>+1.6</td>
</tr>
<tr>
<td>India (Fan)</td>
<td>+1.8</td>
<td>+1.1</td>
</tr>
<tr>
<td>U.S.S.R.</td>
<td>+3.6</td>
<td>-0.2</td>
</tr>
<tr>
<td>Fed. Rep. Germany (Graw)</td>
<td>+0.3</td>
<td>0.0</td>
</tr>
</tbody>
</table>

In India, the chronometer type is used at all stations north of Lat. 22° including Bombay, the fan type at the other stations south of 22°N.
Another example of such small variability of the average summer temperatures can be given by Lhasa (29.7°N), on the southern part of the Tibetan plateau:

<table>
<thead>
<tr>
<th></th>
<th>500</th>
<th>400</th>
<th>300</th>
<th>200 mb</th>
</tr>
</thead>
<tbody>
<tr>
<td>July - August 55</td>
<td>-0.8</td>
<td>-10.7</td>
<td>-23.8</td>
<td>-45.2°C</td>
</tr>
<tr>
<td>56</td>
<td>-0.6</td>
<td>-10.6</td>
<td>-23.2</td>
<td>-44.4</td>
</tr>
<tr>
<td>57</td>
<td>-1.2</td>
<td>-11.3</td>
<td>-24.0</td>
<td>-45.8</td>
</tr>
<tr>
<td>58</td>
<td>-1.0</td>
<td>-10.6</td>
<td>-23.6</td>
<td>-45.4</td>
</tr>
<tr>
<td>59</td>
<td>-0.7</td>
<td>-10.7</td>
<td>-24.0</td>
<td>-46.3</td>
</tr>
<tr>
<td>60</td>
<td>-1.3</td>
<td>-10.8</td>
<td>-24.0</td>
<td>-46.6</td>
</tr>
</tbody>
</table>

Neighbouring stations, such as Heiho and Changtu, yield similar results, as well as do Douala (Cameroons) or Aoulef (Sahara) for example. If we consider such small temperature changes from year to year as typical for the behaviour of the tropical summer atmosphere above southern Asia, we become very doubtful about the consistency of the following averages for July/August (which are admittedly not homogeneous in time) from 4 stations at nearly equal latitude:
<table>
<thead>
<tr>
<th>Location</th>
<th>850</th>
<th>700</th>
<th>500</th>
<th>300</th>
<th>200</th>
<th>150 mb</th>
<th>Period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bombay (17.8°N)</td>
<td>19.1</td>
<td>11.8</td>
<td>-1.5</td>
<td>-23.3</td>
<td>-43.3</td>
<td>-56.2</td>
<td>1957-62</td>
</tr>
<tr>
<td>Vishakhapatnam (17.7°N)</td>
<td>20.4</td>
<td>11.5</td>
<td>-3.7</td>
<td>-29.0</td>
<td>-51.4</td>
<td>-66.8</td>
<td>1957-62</td>
</tr>
<tr>
<td>Rangoon (16.9°N)</td>
<td>17.9</td>
<td>9.8</td>
<td>-4.2</td>
<td>-27.6</td>
<td>-49.2</td>
<td>-62.7</td>
<td>4 months</td>
</tr>
<tr>
<td>Chiengmai (18.8°N)</td>
<td>17.6</td>
<td>9.4</td>
<td>-5.1</td>
<td>-29.8</td>
<td>-52.1</td>
<td>-66.2</td>
<td>1955-57</td>
</tr>
</tbody>
</table>

Similar horizontal and time inconsistencies can be found also in other tropical areas, especially above Africa, where at least five different types of radiosondes have been used. It is therefore by no means surprising, that geostrophic winds derived from such data are to a large extent unrealistic. Above India, the recent radiosonde data (Fig. 5) show a distribution of westerly and easterly components rather similar to that of actual winds (Fig. 6). However the core of the TBE is shifted from about Lat. 15°N to 22°N, with a greatly exaggerated speed of 120 ms⁻¹ at the 100 mb-level. The relative minimum of geostrophic winds at 12-16°N is completely unrealistic. The possible occurrence of a second core south of 10°N cannot be excluded, since Koteswaram (20) has presented some supporting evidence based on data from Gan and Colombo. A second cross-section derived from all nighttime registering balloon ascents between 1927 and 1943 (Fig. 6, upper part) is hampered by the fact that the number of these registering balloons was rather small and their distribution in area and time quite inhomogenous. Therefore the geostrophic zonal wind components are similarly inconsistent with the actual zonal
wind pattern.

3.2 Meridional gradients and cross-sections

Having in mind these inconsistencies of the temperature distribution, we ought to be very cautious in discussing the observed meridional cross-sections and gradients. In all cases where different radiosondes types are used, any quantitative evaluation of horizontal gradients is subject to instrumental errors; in some cases even the sign of such horizontal differences remains doubtful. Nevertheless, in our case we obtain a general qualitative coincidence between the direction of the zonal flow in the TEJ-region and the accompanying meridional temperature gradient. According to the well-known thermal wind equation, any easterly flow increasing with height ought to be correlated with poleward increasing temperatures. If we disregard the layers below 700 mbs (due to local influences) and above 200 mbs (due to increased radiational errors of radiosondes), we obtain the following mean meridional temperature gradients (positive for temperature rising towards north):

<table>
<thead>
<tr>
<th>Meridional temperature gradient S-N, July-August</th>
</tr>
</thead>
<tbody>
<tr>
<td>Longitude</td>
</tr>
<tr>
<td>-----------</td>
</tr>
<tr>
<td>100°E</td>
</tr>
<tr>
<td>78°E</td>
</tr>
<tr>
<td>48°E</td>
</tr>
<tr>
<td>32°E</td>
</tr>
<tr>
<td>5°E</td>
</tr>
</tbody>
</table>
We may add a few averages for the layer 700-200 mbs:

100°E Chiangmai - Songkhla \( (18.8-7.2^\circ N) \) \( 1.3^\circ/1000 \text{ km} \)
92°E Gauhati - Port Blair \( (26.2-11.7^\circ N) \) \( 3.2^\circ/1000 \text{ km} \)
78°E New Delhi - Nagpur/Veraval \( (28.6-21.0^\circ N) \) \( 3.5^\circ/1000 \text{ km} \)
78°E Vishakhapatnam - Trivandrum \( (17.7-8.5^\circ N) \) \( 1.8^\circ/1000 \text{ km} \)

From these values, those along 100°E, 92°E and 5°E seem to be least biased and most representative. From all available temperature data (uncorrected) we selected a number of series for meridional cross-sections. The section along 100°E is certainly one of the most instructive (Tab. 3 a); unfortunately the strongest gradient occurs between the Chinese and the Thai network with an unknown instrumental difference. In the section along 78°E (Tab. 3 b) we use for comparison two Tibetan stations which agree fairly well with New Delhi. However the thermal gradient between the stations along 26°N (Jodhpur, Allahabad) and New Delhi is certainly not consistent with the prevailing easterly flow (Fig. 3); the data from Jodhpur and Allahabad must be considered as (relatively) too warm. The strong temperature gradient between 17° and 22°N seems much too large, as shown by the difference between geostrophic and actual wind distribution (Fig. 5 and 6). It is probably exaggerated due to the use of two different radiosonde types in northern and southern India. The atmosphere over the southern tip of the peninsula coincides well with the equatorial atmosphere above the Pacific, as derived from (33). Using this average of Pacific stations \( (7-10^\circ N) \) as a reference atmosphere, we can construct with these data (including three further stations in northern Tibet and the Tarim Basin and Singapore as an equatorial station) a meridio-
nal cross-section of temperature deviations (Fig. 5). The exaggeration of the meridional gradient in the upper troposphere at Lat. 20-23° is easily seen; from the zonal cross-section along 25°N (Tab. 4 a) we may estimate, that the positive temperature deviation above Jodhpur and Allahabad at Lat. 26°N is perhaps 3-4°C too high. A meridional gradient of 8°C at 200 mb between Allahabad and Trivandrum seems to be more consistent with the wind field than the observed values; this gradient should be concentrated somewhat north of Lat. 15°N.

The meridional cross-section along the Nile valley (Tab. 3 c) seems little affected by instrumental errors; here the poleward increase of temperature reaches 7°C at 150 mbs. The meridional temperature sections above Western and Central Africa (Tab. 3 d) use (with exception of Leopoldville) only French radiosondes and are therefore most homogeneous; the meridional increase in the layer 150-200 mb is here only 6°C. The complete absence of a meridional gradient at 500 mb is surprising.

3.3 Remarks on the upper air climatology of Tibet during summer

The persistent occurrence of the TEJ during the summer season is certainly related to the uniquely high temperatures and height of the isobaric surfaces (up to 100 mb) above the Tibetan plateau and its southeastern fringes. These temperatures produce, with remarkable regularity during each summer, a strong baroclinic field in the middle and upper troposphere coinciding with the strong and persistent easterly flow of the TEJ. Comparative studies of the
upper air climatology have been started since aerological data from the extended network of the Chinese Peoples Republic has become available. These data include information from at least six radiosonde stations on the Tibetan plateau, at altitudes between 2800 and 4200 m above sea-level (Tab. 5); for comparison two stations in the area of the meridional gorges (north of Burma and east of Assam) are included. West of 90°E no data are available. Unfortunately winds are only measured by optical methods, thus strongly biased by a selection of fine weather situations; at Lhasa, 500 mb winds from the SW-sector prevail during each summer.

The temperature data from early morning ascents (00 h GMT = 06 h local time) reveal a nearly homogenous distribution in the southeastern part of the highlands (Heiho, Lhasa, Changtu). From the discussion in chapter 3.2 we may conclude, that the similarly high temperatures of the 12 h GMT ascents above northern India (17 h local time) are perhaps slightly too high in comparison. Unfortunately no direct radiosonde comparison exists. The use of the wind field (Fig. 3) and the thermal wind equation seems to indicate that the highest temperatures should be expected right above the Himalayas, i.e. on the subtropical anticyclonic belt forming the transition zone between tropical easterlies and extratropical westerlies.

It should be stressed that average values of 0°C to -1°C, as observed at 500 mb above southern Tibet and (perhaps) northernmost India during summer, are unique above the whole globe (7, 9). The same is true for the height (geopotential) of the high-tropospheric isobaric levels, especially for the 100 mb-level (21).
The mean summer values at 500 mb above the central Sahara (Tab. 3d) are \(-6^\circ C\) to \(-7^\circ C\), nearly the same as above the equatorial Pacific (annually) or above the hottest part of North America. From Egypt to the Persian Gulf, mean values of \(-3^\circ C\) to \(-4^\circ C\) are also lower than those above southern Tibet (Tab. 4 b).

The physical reason for such uniquely high temperatures and similarly for the maximum of potential energy are twofold (12): a) The transfer of sensible heat from the elevated surface of the Tibetan plateau, with an average height of about 4500 m and an expanse of nearly \(2 \cdot 10^6 \text{ km}^2\), together with that from the adjacent mountain areas from eastern Turkey and Iran to western China. Direct measurements of the radiation and heat balance on an arid high plateau have been given by Zuev (42), comparing data from Lake Kara-Kul (Central Pamir, 3990 m) and of Repetek (Karakum desert, ca. 200 m), both at the same latitude of 39°N. The results are (in Langley/d = gcal cm\(^{-2}\)/24 h, measured during August):

<table>
<thead>
<tr>
<th>Effective incoming radiation ((S+H)(1-a))</th>
<th>Pamir</th>
<th>Karakum</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ly/d</td>
<td>546</td>
<td>525</td>
</tr>
<tr>
<td>&quot; outgoing</td>
<td>E - G</td>
<td>-305</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-244</td>
</tr>
<tr>
<td>Net radiation</td>
<td>241</td>
<td>281</td>
</tr>
<tr>
<td>Transfer of sensible heat into the air</td>
<td>240</td>
<td>274</td>
</tr>
<tr>
<td>&quot; latenat heat</td>
<td>+18</td>
<td>0</td>
</tr>
<tr>
<td>&quot; heat into the soil</td>
<td>-17</td>
<td>7</td>
</tr>
</tbody>
</table>

The following notations are used: \(S\) = direct solar radiation, \(H\) = diffuse sky radiation, \(a\) = albedo of the earth's surface, \(E\) = long-wave radiation of the surface, \(G\) = long-wave radiation of the atmosphere. According to these data the net radiation and the transfer of sensible heat into the atmosphere are nearly the
same in both places; thus the elevation of the heat source together with strong convective activity is responsible for higher summer temperatures of the middle layers (600 – 300 mbs).

This effect has been demonstrated in a numerical model computation by Murakami (24), who added a heat source (of 475 Ly/d) to the average 300 mb-flow above the Tibetan area during May and obtained after a 72 hour-interval an anticyclonic cell over the southeastern part of this area, almost exactly at the same place as shown by actual data.

b) In addition to this, the release of latent heat of condensation in the areas of high orographically induced rainfall is another powerful heat source. If we consider an area with an average rainfall of 300 mm/month (1 g/cm$^2$ d), we obtain an average release of 600 Langley/d, i.e. about twice as much energy as furnished by the net radiation. In fact there exist vast areas along the Himalayas with average rainfall in the order of 500 mm per month during the monsoon season (mid-June to early September). Here the release of latent heat is greater than that produced by the global radiation (S+H) in an arid subtropical area during summer (about 800 Ly/d).

It is well known, that in the NE of the Indo-Pakistan subcontinent (Bengal, Assam) and in the adjacent parts of Burma the rainy season starts already during spring, i.e. about 2 months before the reversal of the tropospheric wind pattern. This is caused by the coincidence between a flat orographically induced upper trough in the subtropical westerly jet south of the Himalayas near 90°E and a low-level moist southerly flow produced by the shallow heat-lows above the interior of the peninsula which gradually move
northward during April and May into their final position over the Punjab (12, 26).

The areas of maximum rainfall over the eastern portion of the Himalayas and the meridional ranges of Upper Burma coincide quite well with the center of the anticyclonic cell. This center is situated, after the first Chinese maps (34), at 30°N, 96°E, and after Murakami's computations (24) near 31°N, 98°E i.e. over the above-mentioned area of homogeneous temperatures in SE-Tibet, slightly north of the area of maximum rainfall. The high relative humidity above these stations (Tab. 5) support the coincidence between rainfall (i.e. lifting) and anticyclonic conditions aloft, as suggested by Riehl (31) in a case study on the production of kinetic energy by the release of latent heat above the Caribbean. This effect of persistent convective rainfall may also be responsible for the remarkably small standard deviation of temperature at 500 mb above Tengchung (Tab. 6), when compared with Lhasa or Heiho, or with Bahrain at the same latitude.
4. DYNAMICS OF THE TROPICAL EASTERLY JET

4.1 Cross-circulation in a jet-stream

It is well-known, that the average ageostrophic cross-circulation in the entrance region of a jet-stream can be described as a direct cell, with lifting warm air and sinking cool air, while in the exit region the prevailing cross-circulation acts as an indirect cell, with lifting cool air and sinking warm air. While in the entrance region where the average flow is accelerated potential energy is converted into kinetic energy, the reverse is true in the exit region with a decelerating flow. This can be most easily understood if we take into account that in the entrance region the actual wind \( v \) is generally smaller than the geostrophic wind \( v_g \). During the process of adaptation between wind field and pressure field in a more or less zonal flow, the imbalance between the pressure gradient \( \frac{\partial p}{\partial y} \) and the Coriolis force \( C = 2\Omega \sin \gamma \cdot v \) tends to deviate every particle with actual velocity \( v \) into an ageostrophic direction: towards low pressure in areas with accelerating flow. This ageostrophic component of the flow causes in a thermodynamic sense, a work-producing (solenoidal) circulation at the entrance region. A schematic picture of the vertical and the ageostrophic cross-components in the exit region of a westerly jet together with a theoretical discussion, which shall be omitted here, has been given by Newton (cf. Lit. 29, Fig. 4, 221, 3, p. 161).
In the TEJ-region the orientation of the cross-sections is reversed compared with a jet in the westerlies. Here the warm anticyclonic side of the jet is situated on its northern flank, the cool low pressure side on the equatorial flank. The entrance area extends from the Pacific at about 150°E to the area of maximum velocity near 80°E, while the even larger exit area extends from there to the Atlantic near 20°W. If our description of the average vertical and cross-circulation near a jet-stream is correct, we may apply it to the TEJ, and we expect then, at the level of 200-100 mbs, in the entrance region east of the centre of the jet an ageostrophic component (perpendicular to the jet core) from north (Fig. 7).

In this chapter we will try to give a realistic quantitative estimate (11) of the cross-circulation and the divergence of the flow in the area of the TEJ. This should be based, as far as possible on homogeneous actual wind observations. We will use the following notation:

\[ u, v = \text{zonal and meridional components of the time-} \]
\[ \text{averaged resultant wind } \mathbf{v} \text{ at any level; easterly} \]
\[ \text{and southerly components positive,} \]
\[ \bar{u}, \bar{v} = \text{averages of } u, v \text{ at an individual station and} \]
\[ \text{for the layer 500-100 mb, averaged with respect to} \]
\[ \text{pressure } \left( \bar{v} = \frac{1}{n} \int_{500 \text{ mb}}^{100 \text{ mb}} v \, dp \right), \]
\[ \bar{u}, \bar{v} = \text{averages of } u, v \text{ for a group of stations}, \]
\[ \bar{\bar{u}}, \bar{\bar{v}} = \text{averages of } \bar{u}, \bar{v} \text{ for a group of stations}, \]
\[ \bar{\alpha} = \text{direction of the average flow } [\bar{u}, \bar{v}] \text{ at an} \]
\[ \text{individual station.} \]
\[ \vec{a} = \text{direction of the average flow} \begin{bmatrix} \vec{u}, \vec{v} \end{bmatrix}. \]

The vertical integration is extended from 500 mb to 100 mb, equivalent to the layer between about 6 and 16 kms, assuming that the 500 mb-level may represent approximately a level of non-divergence or at least of minimum divergence.

The average horizontal divergence of the layer 500/100 mb over an area is given by

\[ \text{div}_2 \vec{V} = \frac{\partial \vec{u}}{\partial x} + \frac{\partial \vec{v}}{\partial y} \]

4.2 Cross-circulation in the exit region of the TEJ (Africa)

In the large exit region of the TEJ we concentrate our efforts on the comparatively good data above the continent of Africa, where the jet axis deviates slightly from a zonal direction. Defining a line connecting all wind maxima in each meridional cross-section, regardless of their individual heights, as the true jet core axis, we find that between 65°E and 5°W this core axis extends in an average direction from 83° to 263°, i.e. with a deviation of 7° from a latitude circle.

In that area most upper wind data show a weak southerly component, which is more stressed if we use a natural coordinate system with an abscissa parallel to the jet core axis, and an ordinate perpendicular to it. Quantities referred to this natural system will be denoted with the index \( p (u_p, v_p \text{ etc.}) \).

In order to obtain reliable figures from a homogeneous, complete set of data, we restrict our computations to two groups of
stations - using the period July-August 1957-60 (8 months) -

I: N (north of the jet core axis): Khartum, Kano, Niamey, Dakar;

II: S (south of the jet core axis): Aden, Douala, Lagos, Abidjan.

The average latitude of group N is 13.9°N, of group S 7.2°N. Unfortunately wind statistics of other stations (Malakal, Ft. Lamy, Coquihatville, Bangui, Ngaoundere and S. Tomé) with only incomplete data had to be omitted; they support - at least qualitatively - the results obtained here.

The results of the evaluation of these upper wind data (cf. also Fig. 4) are as follows:

<table>
<thead>
<tr>
<th>100</th>
<th>150</th>
<th>200</th>
<th>300</th>
<th>500</th>
<th>100/500 mb</th>
</tr>
</thead>
<tbody>
<tr>
<td>N</td>
<td>$\bar{u}$ = 35</td>
<td>36</td>
<td>30</td>
<td>18</td>
<td>15 kt</td>
</tr>
<tr>
<td>S</td>
<td>$\bar{u}$ = 32</td>
<td>35</td>
<td>38</td>
<td>24</td>
<td>5 kt</td>
</tr>
<tr>
<td>N</td>
<td>$\bar{v}$ = 4.1</td>
<td>4.2</td>
<td>3.5</td>
<td>0.1</td>
<td>0.6 kt</td>
</tr>
<tr>
<td>S</td>
<td>$\bar{v}$ = 2.9</td>
<td>2.1</td>
<td>-3.6</td>
<td>-3.1</td>
<td>-3.7 kt</td>
</tr>
</tbody>
</table>

In this case the resultant wind direction $\bar{\alpha}$ is for N = 95°, for S = 83°, the latter value exactly coinciding with the direction of the jet core axis (cf. chapter 4.3). For the meridional components of the divergence we obtain:

$$\frac{\partial \bar{v}}{\partial y} = 2.0 \text{ ms}^{-1}/6.7^0 \text{ Lat} = 2.0 \text{ ms}^{-1}/745 \text{ km}$$

$$= 2.65 \cdot 10^{-5} \text{s}^{-1}$$
The zonal divergence term can be estimated from the deceleration between Aden ($u = 15.8 \text{ ms}^{-1}$) and Abidjan ($u' = 11.3 \text{ ms}^{-1}$):

$$\frac{\partial \vec{u}}{\partial x} = -4.5 \text{ ms}^{-1}/5500 \text{ km} = -0.83 \cdot 10^{-6} \text{ s}^{-1}$$

Adding both terms – meridional divergence and zonal compression – we obtain the horizontal divergence of the layer 500-100 mb:

$$\text{div}_2 \vec{v} = \frac{\partial \vec{u}}{\partial x} + \frac{\partial \vec{v}}{\partial y} = 1.82 \cdot 10^{-6} \text{ s}^{-1}$$

The meridional outflow produced by this horizontal divergence can be easily estimated, assuming a meridional section with a width of 1 cm extending over the average distance of 6.7°Lat. = 745 km between the two groups N and S, i.e. above an area $F = 7.45 \cdot 10^7 \text{ cm}^2$.

Since the layer 500 - 100 mb contains nearly 0.4 of the total mass $M$ of a column of the atmosphere with a weight $W$ of 1010 g/cm², the meridional outflow $M_y$ is

$$M_y = 0.4 \cdot W \cdot F \cdot \text{div}_2 \vec{v} \quad \text{[g s}^{-1}]$$

Under stationary conditions, this mass outflow can be balanced by a net inflow in the lower troposphere below 500 mbs and in the stratosphere above 100 mbs. Here we assume the stratospheric transport to be zero, and we obtain a meridional circulation with divergence aloft and convergence below and with a vertical mass transport $M_z$ through the 500 mb-level produced by a vertical wind component $\vec{w}_{500}$. If at 500 mb the density $\rho_{500} = 0.65 \cdot 10^{-3} \text{ g cm}^{-3}$,
then

\[ M_z = \bar{w}_{500} \cdot F \cdot \Phi_{500} \quad [\text{gs}^{-1}] \]

Assuming stationary conditions, \( M_z = M_y \), we obtain

\[ \bar{w}_{500} = \frac{0.4 \cdot W \cdot \text{div} \varphi}{500} = 1.13 \text{ cm s}^{-1}. \]

Thus in our model the horizontal outflow at the layer 500-100 mb is balanced by an equivalent inflow below 500 mb and an area-averaged uplift of 1.13 cm s\(^{-2}\) for the large area between Lat. 7.2\(^\circ\)N and 13.9\(^\circ\)N and Long. 45\(^\circ\)E and 5\(^\circ\)E, covering about 4.1 \cdot 10^6 \text{ km}^2 (cf. Fig. 8A). As this area coincides with the tropical summer rain belt and the zone of maximum convective activity, such a relatively large average uplift of about 1 km per day seems to be reasonable.

The coincidence of the direction of the jet core axis and of the average direction of flow at the station group S (\( \bar{y} = 7.2^\circ\)N) in the layer 500-100 mb suggests a second model, where the axes of our terrestrial coordinate system are slightly rotated by an angle of 7\(^\circ\). In this model the flow perpendicular to the southern boundary (near 7\(^\circ\)N) \( m_p \) is zero, and we can assume a closed system between two solid vertical walls, both running parallel from 83\(^\circ\) to 263\(^\circ\). The southern wall coincides with our station group S and crosses Long. 20\(^\circ\)E at 7.2\(^\circ\)N. The northern wall coincides approximately with the subtropical divergence axis, where in the layer 500-100 mb the frequency of easterly and westerly winds is equal, and its position is fixed by the intersection of 25\(^\circ\)N and 20\(^\circ\)E.
The total area is divided by a (permeable) middle wall running parallel to the outer walls through 13.9°N, 20°E into the southern area I, coinciding with the TEJ-core, and the larger northern area II. In this closed system the outflow from the TEJ-area I in the layer 500-100 mb is equal to the inflow from the north below 500 mb (Fig. 8 B). For a closed circulation cell we have to assume, in the larger northern area II, an average sinking motion through the 500 mb level. Since the meridional diameter of I (II) is 6.7 (11.1)° latitude, we obtain

\[ w_{500}^{(II)} = - \frac{6.7}{11.1} w_{500}^{(I)} = -0.68 \text{ cm s}^{-1} \]

If our model assumptions are sufficiently realistic, we obtain for area II, covering some 6.7 \times 10^6 \text{ km}^2 between about Lat. 14°N and 25°N, a superimposed sinking motion in the order of 600 m/d. Due to the model assumptions and to the inadequacy of the aerological network, the numerical values of the vertical and quasi-meridional components should be understood with an uncertainty of ± 10 – 20 percent.

During July–August our area II coincides quite well with the dry subsidence zone on both flanks of the Intertropical Convergence Zone (ITC) or of the Equatorial Pressure Trough, together with a narrow belt of shallow cumulus clouds 200-300 kms south of it, i.e. with a belt of suppressed convective activity. In this belt, the frequency of convective showers is near zero in the ITC and north of its average position near 18°N, and it is weak (5–8 days
a month) and inefficient in a belt 200-300 kms wide south of the ITC (41).

The strange asymmetric distribution of convective activity and weather on both sides of the ITC has been ascribed hitherto, in the publications of British and French meteorologists (23, 41), to the conservative properties of air masses of different origin. Since the vertically averaged water-vapour transport in this belt is, to a large extent, directed from E to W, i.e. back from the continent to the ocean, due to the shallowness of the lower southwest "monsoon", this air mass hypothesis is hardly satisfactory. This has already been pointed out by several authors, who stressed the large difference in stability and humidity between the air masses reaching the central Sudan from the Atlantic, those from the Mediterranean or those from the Indian Ocean, all travelling nearly equal distances over the continent. In an earlier publication (8) the author has shown that a similar asymmetric distribution of the divergence and the vertical component of wind (derived from the continuity equation) can be found along an oceanic cross-section over the Atlantic from the wind-field alone. Here we obtained from surface and pilot balloon data above the Atlantic (20-30°W), average vertical components of + 1.0 - 1.5 cm s"1 in a distance of 3-6°Lat. south of the ITC, for an altitude of 4-5 kms, together with sinking motion above 4-6 kms north of the ITC [Fig. 9]. This result interprets the asymmetric distribution of convective activity (and air-mass properties) in terms of atmospheric dynamics, yet without giving a physical explanation for the origin of the asymmetry. The above-mentioned result may con-
tribute to such a physical explanation, at least during the northern summer, where the asymmetry is most pronounced. By these investigations the hypothesis is supported, that the extraordinary intensity and persistence of the summer-time subtropical anticyclone above North Africa and Southwestern Asia is strengthened by the ageostrophic flow in the large exit region of the TEJ, extending from 75°E to about 15°W. This is also true for the area of prevailing strong subsidence on both sides of the equatorial pressure trough over the Sahara, Arabia and the regions of the Persian Gulf up to Sind and Baluchistan, where the dynamically forced sinking motion tends to suppress any large-scale convective activity. In this whole exit area the very slow gradual deceleration of the jet core results in widespread convergence aloft and sinking motions on the warm side. In this vast area we observe an indirect meridional circulation cell with sinking motions at the subtropical warm cell on the northern flank and rising motions on the relatively cool equatorial side, i.e. an Anti-Hadley cell above about one quarter of the earth’s circumference. The role of this indirect cell within the general atmospheric circulation will be discussed in chapter 5.

4.3 Cross-circulation in the entrance region (Southeast Asia)

In the entrance region above southeastern Asia the isotach analysis of the TEJ seems to reveal a more complicated pattern, perhaps consisting of two branches. This is also likely in the section along 100°E, where four stations between 1°N and 19°N
provide a fairly reliable cross-section which, however, is not quite homogeneous with respect to time. Here the most reliable data for July/August are:

Resultant wind vectors $\vec{v}$ of the layer 500-100 mbs (knots)

<table>
<thead>
<tr>
<th>Station</th>
<th>Lat.</th>
<th>Long.</th>
<th>$\vec{u}$</th>
<th>$\vec{v}$</th>
<th>$\vec{z}$</th>
<th>$\vec{v_r}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chiengmai</td>
<td>18.8°N</td>
<td>99.0°E</td>
<td>-20.3</td>
<td>-4.69</td>
<td>77°</td>
<td>20.8</td>
</tr>
<tr>
<td>Bangkok</td>
<td>13.7°</td>
<td>100.5°</td>
<td>-17.8</td>
<td>-4.24</td>
<td>77°</td>
<td>18.3</td>
</tr>
<tr>
<td>Songkhla</td>
<td>7.2°</td>
<td>100.6°</td>
<td>-21.7</td>
<td>-4.78</td>
<td>78°</td>
<td>22.2</td>
</tr>
<tr>
<td>Singapore</td>
<td>1.3°</td>
<td>103.0°</td>
<td>-23.4</td>
<td>-3.19</td>
<td>82°</td>
<td>23.6</td>
</tr>
<tr>
<td>Guam</td>
<td>13.6°</td>
<td>144.9°</td>
<td>-7.05</td>
<td>-1.85</td>
<td>75°</td>
<td>7.3</td>
</tr>
<tr>
<td>Clark Field</td>
<td>15.2°</td>
<td>120.0°</td>
<td>-11.9</td>
<td>-1.40</td>
<td>83°</td>
<td>12.0</td>
</tr>
<tr>
<td>Hongkong</td>
<td>22.3°</td>
<td>114.2°</td>
<td>-12.9</td>
<td>-2.78</td>
<td>78°</td>
<td>13.2</td>
</tr>
<tr>
<td>Madras</td>
<td>13.0°</td>
<td>80.2°</td>
<td>-22.0</td>
<td>-2.38</td>
<td>84°</td>
<td>22.1</td>
</tr>
<tr>
<td>Vishakhapatnam</td>
<td>17.7°</td>
<td>83.2°</td>
<td>-21.5</td>
<td>-4.18</td>
<td>79°</td>
<td>21.9</td>
</tr>
</tbody>
</table>

From these data we are entitled to suggest the probable existence of two cores within the total cross-section, the northern one near 16°N and 100 mbs, the second one near 5°N and 150 mbs. Under this assumption we may calculate two independent values for $\frac{\partial \vec{v}}{\partial y}$ in the layer 500-100 mb:

- Chiengmai-Bangkok: $\frac{\partial \vec{v}}{\partial y} = -0.41 \cdot 10^{-6} \text{ s}^{-1}$
- Songkhla-Singapore: $\frac{\partial \vec{v}}{\partial y} = -1.25 \cdot 10^{-6} \text{ s}^{-1}$

average: $\frac{\partial \vec{v}}{\partial y} = -0.83 \cdot 10^{-6} \text{ s}^{-1}$

These values, however, seem to be less representative than an
average taken over the large distance from Chiengmai to Singapore = \(-0.40 \cdot 10^{-6} \text{ s}^{-1}\).

A realistic estimate of \(\partial \vec{u} / \partial x\) can be derived from a zonal cross-section Clark Field (Luzon) - Bangkok - Madras:

- Guam-Clark Field: \(\partial \vec{u} / \partial x = +0.93 \cdot 10^{-6} \text{ s}^{-1}\)
- Clark Field-Bangkok = +1.45 " "
- Bangkok - Madras = +0.99 " "
- Guam - Madras = +1.10 \cdot 10^{-6} \text{ s}^{-1}

For comparison we compute also \(\partial \vec{u} / \partial x\) from a zonal section along 18°-22°N:

- Hongkong - Chiengmai = +2.30 \cdot 10^{-6} \text{ s}^{-1}
- Chiengmai - Vishakhapatnam = +0.36 " "
- Hongkong - Vishakhapatnam = +1.32 \cdot 10^{-6} \text{ s}^{-1}

Averaging these components for the simple model, we obtain a fairly representative value of the horizontal divergence of the layer 500-100 mb in the entrance region:

\[
\text{div}_2 \vec{V} = \partial \vec{u} / \partial x + \partial \vec{v} / \partial y
+1.10 -0.40 = +0.70 \cdot 10^{-6} \text{ s}^{-1}
\]

In contrast to the exit region \(\partial \vec{V} / \partial y\) is comparatively small, so that \(\text{div}_2 \vec{V}\) once again is positive. Disregarding the stratosphere, this must be balanced in the layer below 500 mbs by convergent (southwesterly) flow which dominates - according to the fine atlas of upper winds over Southeast Asia [5] - up to the 700 mb-level over the whole area between Lat. 18° and 29°N.
Long. 90° and 120°E. Due to the general northerly component above 500 mbs, the outflow from the entrance region of the TEJ is directed to the south. In a latitudinal area extending from Luzon to Madras and between Lat. 18.8°N and 1.3°N - i.e. F = 8.15 \cdot 10^6 \text{ km}^2 - this value of \text{div}_2 \mathbf{\nabla} necessitates an area-averaged lifting motion at the 500 mb-level of \( \bar{\mathbf{w}}_{500} = + 0.44 \text{ cm s}^{-1} \). Taking into account the rainfall pattern during this season, it seems most likely that the area of large-scale lifting motion extends much farther to the north, at least to about 33°N.

In this simple model, as well as in the above-mentioned two-core model, the occurrence of a region with prevailing lifting motion on the northern flank of the TEJ-system - i.e. in the latitudinal zone 20-30°N - is established by the frequent heavy summer rains in this mountainous area. On the southern flank, a few oceanic or low-level areas near the equator - such as the Java Sea between Java, Borneo and Malaya - show indications of prevailing subsidence. In our simple one-core model, we may deduce the marked seasonal dryness of the atmosphere above the Indonesian waters at 5-10°S - indicated also by the low humidity above the 700 mb level at Djakarta, as discussed by Trewartha (36), and in marked contrast to the same latitudes over the Indian ocean - from this cross-contour circulation. If our data along 100°E are representative, there is little space for large-scale subsidence in the TEJ-region north of the equator.

Strong evidence for a direct (solenoidal) circulation over the Bengal Sea - as suggested by Koteswaram (19) - is available. According to the maritime data in the meridional strip 82-88°E
(9), the frequency of rain during July and August increases from 2-5 % of all ship observations at Lat. 5-10°N to more than 20 % at Lat. 18-22°N. A similar phenomenon can be demonstrated along the west coast of the Deccan (27), where the areas of maximum (orographic) rain are situated near Lat. 14° and 18°N, while south of 12°N the rain amounts are definitely smaller, in spite of the fact that the highest peaks occur in that southern area. A maximum near 15°N is also very prominent in the rain frequency data of the Arabian Sea (62-68°E), together with a minimum near 10-12°N (Lit.9, Figs. 4 a and 6).

4.4 The reversal of the cross-circulation

In the case of such a quasipermanent jet stream we have to consider the question: what mechanism produces a reversal of the ageostrophic circulation from the entrance to the exit region? At first it should be stressed, that in our case the average non-ageostrophic cross-component of the flow is about 10 % of the more or less zonal flow; i.e. a particle travelling with the flow will cross the core of the jet at a comparatively small angle (of the order 5°-10°). But even so; the marked transition from acceleration to a deceleration cannot be attributed to inertial forces or to internal friction. In this transition zone the TEJ crosses near the area of its maximum speed two more or less meridional belts of orographically fixed high convective activity and monsoonal rains; both are situated along the west-facing, mountainous coasts of the large-scale peninsula of Burma-Thailand-Malaya
(93-98°E) and of the Deccan plateau of India (73-77°E), both extending from about Lat. 22°N to at least 8°N, i.e. right across the TEJ-belt. Here the rainfall frequency amounts to 22-30 days a month from the middle of June to early September, and the large intensity of rains (up to 1200-1400 mms/month) indicates very strong vertical exchange processes. This is also demonstrated by the frequent occurrence of a "bright band" in the radar pictures of the monsoon rains from Poona and from airplane observations along both coasts. The cumulonimbus towers along both coasts extend very frequently to above 11 kms, where the average temperature drops below -40° C, thus causing spontaneous freezing of the cloud droplets. In these curtain-like meridional belts of maximum convective activity strong vertical momentum exchange between the westerlies below 500 mbs and the easterlies above is a quite regular phenomenon, which necessarily leads to a deceleration of both opposite currents, together with a reversal of the large-scale cross-component.

On the other hand the relatively weak convective activity above the Arabian Sea contributes - due to the marked speed divergence of the low-level SW-monsoon at this area - to the remarkable constancy of the speed of the TEJ between India and Arabia (Aden). Koteswaram (19) has suggested that the "burst of the monsoon" along the Malabar coast occurs in association with the advance of the TEJ over southern India to the north and with the reversal from direct into indirect vertical circulation.
4.5 Origin of the kinetic energy of the TEJ.

The kinetic energy of the TEJ may be produced by several mechanisms:

a) zonal advection of easterly momentum to the equatorial branch of the TEJ in the entrance region,

b) equatorward displacement of angular momentum in the northern branch of the TEJ in the entrance region,

c) conversion of available potential energy into kinetic energy in the area of increasing meridional pressure gradient southeast of the high-tropospheric Tibetan anticyclone.

The role of a) is most probably negligibly small. The wind systems of the equatorial Pacific in the troposphere are weak, and in the stratosphere the strange biannual periodicity of downward moving rings with opposite (westerly-easterly) motion should be reflected in the strength of the high-tropospheric TEJ. A systematic comparison of the resultant winds in the core of the TEJ above India — including its entrance region above Thailand and its exit region above Africa — does not show sufficient evidence of a contribution from biannual fluctuation (cf. Chapter 2.3).

Since in the area of the subtropical jet at 40-45°N and near the 200 mb-level the westerlies apparently intensify from Central Asia to Japan, it is reasonable to assume that a portion of this flow is diverted above China and finally reversed into the northern branch of the TEJ. In addition to this, it cannot be excluded that the high speed of this northern branch at the 100 mb-level could be derived partly from the stratospheric easterlies, which are very persistent and relatively strong (up to more than 60 knots).
above 50 mb in these subtropical latitudes.

Together with a slow sinking motion above the tropopause, easterly zonal momentum from the stratosphere may enter the upper tropical troposphere and may be increased by equatorward displacement, as indicated by the remarkable frequency of northerly components in the entrance region. At present no direct evidence for this hypothesis is available.

Effect c) seems to be the most powerful, due to the unique intensity of the great Tibetan anticyclone at the 100 mb-level during summer. Having investigated the energy conversions between different wave numbers, S. Teweles [35] concludes, that in the subtropical anticyclone belt the principal conversion is from eddy available potential energy to eddy kinetic energy and takes place in wave numbers 1 and 2. In the main entrance area of the TEJ between 80°E and 120°E (cf. Lit. 35 Fig. 7, 14), both waves act — with a northerly wind component — in the same direction, i.e. produce a strong equatorward transport of eddy kinetic energy at 100 mb (Lit. 35 Fig. 7, 23), converging south of Lat. 20°N. Here it can be assumed, that effects b) and c) act together, due to the unique pressure distribution in that area east of the great anticyclone.

The formation and maintenance of the great Central Asiatic summer anticyclone has been discussed in chapter 3.3; its vital role in the energetics of the TEJ can hardly be overestimated. In order to achieve a deeper understanding of the processes involved, a comparison between the velocity of the TEJ (as correlated with latitude) and the pressure and temperature anomalies above eastern Tibet is suggested.
5. THE ROLE OF THE TROPICAL EASTERLY JET IN THE GENERAL ATMOSPHERIC CIRCULATION

5.1 Hadley and Anti-Hadley circulation

In the area of the TEJ we observe (cf. chapter 4.1), at the same latitude, two opposite meridional cells of the planetary circulation: a direct circulation with rising warm air in the north, sinking cool air on the other side of the equator between about Long. 150°E and 80°E, in contrast to an indirect circulation with sinking warm air and rising cool air between about Long. 70°E and 20°W, i.e. above one fourth of the earth's circumference. If we use the term "meteorological equator" for the belt of maximum temperatures 30°N (above Asia) or 25°N (above Africa), we certainly can describe the direct work-producing circulation in the entrance region as the usual type of Hadley circulation. It thus seems reasonable to define the indirect work-consuming cell in the exit region as an "Anti-Hadley circulation". Discussions with meteorologists from other continents, however, revealed that - compared with the usual latitudinal distribution during winter - we may also define, in a merely kinematical sense, the direct cell above SE-Asia as an Anti-Hadley-regime between subtropical and equatorial latitudes. In order to avoid any possible misunderstanding, the author prefers the first definition, which coincides well with a dynamical and thermodynamical view, correlated with the seasonal displacement of the large-scale temperature, pressure...
and wind belts of the atmosphere above the continently controlled section of the Old World (Africa–Asia–Australia, including the Indian Ocean).

Although there is no special evidence available, it seems likely that part of the high-tropospheric cross-circulation in the TEJ-area is essentially non-geostrophic. The occurrence of fairly strong zonal pressure gradients, however, should be taken into account, at least qualitatively. Due to the uncertainties of the pressure and temperature field aloft, the numerical evaluation of isobaric slopes in the upper troposphere in tropical latitudes is strongly biased by systematic instrumental differences. The geopotential height of the 100 mb-level above Tibet is about 300 m higher than above the Atlantic and the Pacific [21]. This yields a geostrophic meridional component in the order of 4 m s⁻¹ at Lat. 30°. This meridional component is opposite to that in lowest layers. Here the zonal pressure difference between the heat-low center over the Punjab (996 mb) and the anticyclones of the northern Pacific or the Azores (1026 mb each) maintain a steady northerly component from the Mediterranean to the Persian Gulf (Shamal, Etesians) and a definite southerly component from Bengal to the Philippines. It should not be overlooked that from mid-June to early September the Punjab heat-low is the deepest of the northern hemisphere, not only as an average, but also frequently on individual maps. In about the same longitudinal sectors these "monsoonal" low-tropospheric wind systems are balanced by high-tropospheric meridional components in the entrance and exit regions of the TEJ. The heating pattern of the atmosphere above the large
continents of the "Old World" produces, at subtopical and tropical latitudes (0–35°N), two tubes with opposite meridional circulations (solenoidal and anti-solenoidal), where the direction of rotation changes near the Indo-Pakistan subcontinent. A similar but much weaker system of tubular meridional circulations seems to occur above the Caribbean and the eastern Pacific off the west coast of Mexico (cf. chapter 5.2).

This picture is much different from the classical idea (Halley 1686, Woeikof 1874) of a monsoonal circulation between continents and oceans. From here we also gain a deeper insight into the large-scale climatological pattern, especially to the occurrence of wide-spread lifting and raininess in the belt 20–30°N on the northern side of the eastern tube, in contrast to the strong subsidence and aridity (during the summer months) in the same belt on the northern side of the western tube.

5.2 Tropical circulation in the northern hemisphere, during summer
During the northern winter, the quantitative intensity of the Hadley cell averaged over the whole hemisphere has been investigated by Mintz and Lang (22) and by Palmén (25). From their investigations the sinking motion in subtropical latitudes was estimated, in good agreement, at -0.27 cm s⁻¹ or -0.31 cm s⁻¹, respectively. A similar estimate for the northern summer, by Mintz and Lang, yielded only -0.05 cm s⁻¹. This is not only due to the rather crude approximations used in this model. Our investigations show, that above a latitudinal strip of 90–100°Long.
(between 75°E and 20–25°W) the tropospheric circulation between the equator and subtropical latitudes is reversed in comparison to the section between 145°E (Guam) and 75°E. While in the exit region of the TEJ we obtained a sinking motion of \(-0.68\) cm s\(^{-1}\) at subtropical latitudes, we have to expect, in the entrance region, a lifting motion of the same order of magnitude at the same latitudes (cf. chapter 4.3). This view is supported by the large amount of summer rainfall in this region, in remarkable contrast to the subsidence in the arid belt from Rajasthan to the western coast of Africa.

In the American section of our globe, we apparently observe a similar pattern, but much less intense. Alaka (1) has described a single case of an easterly jet above the Caribbean. According to the upper wind data of San Juan (Puerto Rico) and Swan Island we observe, during summer, persistent easterlies of 10–20 knots between 100 and 200 mbs, in most cases increasing with height above 100 mbs. The average 100 mbs maps likewise demonstrate a weak anticyclonic cell above the southwestern United States (New Mexico). The distribution of summer rainfall in the Caribbean, as compared with that of the Pacific off the Mexican coast, seems to indicate a similar checkerboard pattern. Subject to further verification, there seems to be sufficient evidence for a similar pattern in the sector between about 65°W and 130°W, related to a weak but persistent easterly flow in the upper troposphere and to the reversal of zonal gradients along 30–35°N. In contrast to the TEJ above the eastern hemisphere, this easterly flow may be defined as a jet only in rare and exceptional cases.
During the northern summer, the heating processes above the tropical continents and especially above their mountain areas produce in the high troposphere two areas of easterly flow, accelerating east of the position of maximum heating and decelerating west of it. As both anticyclonic cells at the 100 mb-level are about 180° apart, and as the Tibetan cell is much stronger than the American cell, this pattern coincides with a prevalence of the first two wave numbers. The reversal of the cross-contour circulation between the entrance and exit regions of this high-tropospheric easterly flow is obviously responsible for the weakness of the latitudinally averaged values of the meridional and the vertical components of the tropical circulation. These small planetary averages of $\vec{v}$ and $\vec{w}$ are nothing but the residuals of large sections with opposite sign. During summer, the zonal temperature and pressure differences aloft between mountainous continents and oceans are much greater than the meridional differences averaged over the whole globe. The reversal of the tropical meridional circulations, probably repeated twice above Asia-Africa and America, is only a consequence of the "monsoonal" heating pattern; the residual planetary circulation is much weaker than in winter and hardly distinguishable.

5.3 Comparison with southern hemisphere continental sections

Above the southern hemisphere continents, only weak homologies of the TEJ are known during the southern summer (Tab. 1 f). Bond (3) has described two individual easterly jets with wind speeds up to
81 knots at an altitude of 15-16.5 km above northern Australia. The average winds at 200-100 mbs above Darwin during January and February are easterlies, their speed being only about 20 knots. According to all available data, the wind speed increases with height in the stratosphere above 100 mbs. Similar features can be observed above southern Africa (Kasama, Lilongwe, Luanda) and above South America (between Lima and Antofagasta). Above Nairobi (1.3°S) strong easterlies up to 77 knots have been described (6) during December-February as well as from June to August, but not during the transitional seasons. The analysis of African winds by Hofmeyr (16) indicates a core of easterly winds at Lat. 4-8°S and near 200 mb, separated from the stratospheric core of easterlies above 18°S.

In all areas, prevailing easterly winds are observed during January and February in the layer 200 (150) - 100 mbs at Latitude 10-15°S, and they increase with height above 100 mbs. Only in rare individual cases does their speed reach 60-80 knots. In the average, these high-tropospheric easterlies are only the lower extensions of the persistent summer easterlies in the stratosphere. Their role in the general circulation is hardly comparable with that of the TEJ; further investigations are suggested.

5.4 The unique summer aridity of the Sahara, Near and Middle East.

From this point of view our understanding of the enormous arid belt of the Old World can also be improved. This arid belt extends from the Atlantic off the western coast of Africa right into the
heart of the Asiatic continent, i.e. over a zonal section of about 110°Long., but without any counterpart on the eastern side of the Asiatic continent. In our considerations, we exclude the orographically enclosed Tarim basin and the winter-dry Gobi semi-desert; here we shall deal only with the extent of the aridity during the northern summer season. In contrast to classical textbook ideas, the aridity of this vast area during summer is by no means self-evident. This becomes obvious, when we compare the extent of summer rains in this arid belt with that in other continents, including the southern hemisphere.

In the Sahara and Arabia, tropical summer rains extend only, as a regular phenomenon, to about 16°-18°N, and in form of irregular events up to about 22°N. In the belt between 22°N and about 30°N we observe a few scattered rains of tropical origin – or produced by an interaction of tropical low-level disturbances with overlying high-tropospheric troughs in the westerlies – only during May and early June, and then again during September and October. One notable example of this kind are the exceptional showers (up to 40 mms) of August 28, 1944 at the Nile delta. The non-occurrence of such tropical rains during the TEJ-season (mid-June to early September) has to be explained.

In all other continents the summer rains extend much farther pole-ward:

a) In southern Africa the tropical summer rains cover southern Rhodesia and the Kalahari semi-desert; they extend right into the center of South Africa and merge with prevailing summer rains in the belt of westerlies. At Pretoria (26°S) the summer rains are
partly of tropical origin (39); in the meridional cross-section along 32°E (Fig. 10) the southern end of the tropical summer rain belt is well defined at 24°S.

b) The most convincing example is Australia, where the tropical summer rains extend at least to 24-26°S, occasionally to 30°S. The large zonal dimension of Australia (40° Longitude) is certainly smaller than that of northern Africa (68° Long.), but this is hardly sufficient to explain the striking difference in latitude. As a consequence the average rainfall remains at least 100 mms, even in the center of the great Australian desert, while in the broad central belt of the Sahara (including Arabia) the average amount of rainfall remains below 20 mms. The quite different vegetation distribution in both deserts is well-known.

c) On the continent of South America, the merger of tropical summer rains with extratropical summer rains within the westerlies is well established. Therefore a continental desert in subtropical latitudes exists only on the narrow Andean altiplano; here the orographically controlled arid zones on the lee side of the Andes shift from the western slope in the north to the eastern side in the south. On the Andean plateau the tropical summer rains extend to at least Lat. 21°S.

In all southern continents the large poleward extension of tropical summer rains is especially remarkable, because the average position of the meteorological equator is definitely asymmetric and shifted into the northern hemisphere, at about Lat. 5°N.

d) In Mexico and the southwestern United States we also observe a
merger of tropical summer rains with continental summer rains extending right into the center of the Arizona heat low. The mechanism of the summer rains in Arizona has been investigated by Bryson and Lowry (4).

From these comparisons we come to the conclusion that the extreme summer aridity of the dry belt from the western coast of Africa to the Indus, at latitudes 20–30°, is exceptional and asks for better physical explanation, which our concept of a large-scale indirect circulation within the TEJ-system (Chapter 4.3) may yield. During the summer season, the extreme dryness of this area is caused by large-scale subsidence, dynamically forced by the persistent outflow and the southerly cross-component in the exit region of the TEJ. This cross-circulation suppresses nearly completely any convective activity, even in the area of the shallow heat-lows of the ITC with high convergence in the lowest layers.
6. REMARKS ON THE SYNOPTIC BEHAVIOUR
OF THE TEJ ABOVE INDIA

Any investigation on the day-to-day variations of the synoptic pattern of the TEJ above India is seriously hampered by the small frequency of successful ascents into the layer 200-100 mbs. At the stations south of Lat. 22°N only 20% of all ascents reached the 100 mb-level during the years 1956-59. Due to the use of improved balloons the percentage of ascents reaching 100 mbs increased substantially during 1961-62, but is still not comparable to the aerological networks of North America or Europe. Therefore a reliable and sufficiently detailed synoptic routine analysis up to 100 mbs is still difficult over the Indo-Pakistan subcontinent. Under these circumstances we shall try to obtain further information from statistical parameters. As shown in another paper (13), such parameters can be easily interpreted in a synoptic sense.

The most remarkable phenomenon is the high constancy of the TEJ; in its core the steadiness \( q = \frac{v_r}{\bar{v}} \) (\( v_r \) = resultant vector wind speed, \( \bar{v} \) = mean scalar wind speed) usually reaches values between 96 and 100%. This parameter describes only the variability of the wind direction. At the 16.3 km level above Madras we found in a sample of only 105 observations (1956-59) an average wind speed of 69 knots, with individual values ranging between 12 and 120 knots. Wind speeds between 45 and 95 knots are nearly equally frequent, without a clearly defined peak; the curtosis of the sample was unusually high. Wind speeds above 100 knots occur usually in groups
of several days simultaneously at more than one station; such strong wind periods are separated by other periods with generally weak winds. During 1958 and 1959, the following examples of both types have been found:

<table>
<thead>
<tr>
<th>Strong wind periods</th>
<th>Weak wind periods</th>
</tr>
</thead>
<tbody>
<tr>
<td>(Maxima above 100 kts)</td>
<td>(Maxima below 100 kts)</td>
</tr>
<tr>
<td>August 1–6</td>
<td>August 21–28</td>
</tr>
<tr>
<td>1959 July 16–18</td>
<td>1959 June 22–26</td>
</tr>
<tr>
<td>August 11–14</td>
<td>July 22–26</td>
</tr>
</tbody>
</table>

The occurrence of such pulsations of strong and weak wind speeds will be investigated at a later occasion. The highest wind speed during the period 1956–62 was measured above Bombay on July 17 and 18, 1959, with 152 knots each day near 18 k.ms. Even during June (but not before June 10) such high values are observed, e.g. several speeds up to 122 knots at 15–18 kms, on June 22, 1958 and June 16, 1959, at Madras and Trivandrum, respectively. We may also note (without discussion of the reliability) one case of high speeds in the stratosphere, such as on August 10, 1958, with values of 137 and 142 knots resp. at 23–24 k.ms above Nagpur and Veraval. During the period of June 10–August 31, 1958 and 1959 wind speeds above 100 knots at any one of the stations south of Lat. 22°N are observed on 38 % of all days. This value can be biased, however, by any selection of synoptic situations. During June, such high speeds are restricted to latitudes south of 15°N.
A more general parameter representing the persistency of winds is given (13) by the percentage the kinetic eddy energy \( E_E \) represents of the total kinetic energy \( E_T \):

\[
\frac{E_E}{E_T} = \frac{\sigma^2}{\sigma_x^2 + \sigma_y^2}
\]

with \( \sigma^2 = \sigma_x^2 + \sigma_y^2 \) denoting the vector variance of the winds (\( \sigma \) = standard vector deviation). Representative values of \( E_E / E_T \) were obtained by an IBM computer program and are collected in Table 7, which demonstrates the increase of the eddy kinetic energy with distance from the core of the TEJ. The values for stations and heights near the core (Madras 14 km 8-9 °) belong to the lowest values hitherto known, stressing the unique persistency of the TEJ at least compared with all other tropospheric wind systems.

The relationship \( \sigma_x / \sigma_y \) may be interpreted (13) as an indicator for the preferred position of the wind maxima: if \( \sigma_x > \sigma_y \), then the cores of the wind maxima occur along the zonal sections of any wave pattern; i.e. the fluctuations of a (more or less zonal) wind field can be described as pulsatory. In the TEJ-region – as in most parts of the tropical zone – \( \sigma_x \) is significantly larger than \( \sigma_y \) (Table 7); pulsatory motions seem to prevail.

The exceptionally high values of \( \sigma_x / \sigma_y \) at 18 km above Trivandrum (Lat. 8.5°) seem to be influenced by the 26-month-period of equatorial stratospheric winds (cf. 13). Only occasionally are distinct wave patterns observed, as described by Koteswaram (20). The correlation coefficient between the x- and the y-component of
the wind \( \{r_{xy}\} \) is mostly small, which can be interpreted to mean that no predominant tilt of any wave-like motions is observed, and that the meridional transport of zonal or angular momentum remains weak. Only in the vicinity of the core (at 16 km south of 18°N) do significant positive values prevail, thus indicating a southward transport of easterly momentum.

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TABLES

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   a) Southwest Pacific
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5. Average Temperature and Humidity above Tibet and Adjacent Regions, July-August

6. Standard Deviation of Temperature, July-August

7. Statistical Wind Parameters over Indo-Pakistan (1958-62)
### Tab. 1 Resultant Winds (July-August)

#### a) Southwest Pacific

<table>
<thead>
<tr>
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<th>100 mb</th>
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<td>10</td>
<td>33°</td>
</tr>
<tr>
<td>Iwo Jima</td>
<td>1956-59</td>
<td>58°</td>
<td>11</td>
<td>62°</td>
</tr>
<tr>
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<td>10</td>
<td>53°</td>
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<tr>
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<td>13</td>
<td>67°</td>
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<td>69°</td>
<td>17</td>
<td>68°</td>
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<tr>
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#### b) Southeast Asia

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#### c) India and Pakistan

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### f) Southern Hemisphere [January-February]

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1) height 15.3 km

2) height 18.4 km
### Tab. 2 Resultant Winds for Individual Years, July/August

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- direction, V_r = speed of resultant wind (knots), q = steadiness, n = number of ascents
### Tab. 3a Meridional Temperature Cross-Section along 100°E

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### Tab. 3b Meridional Temperature Cross-Section along 78°E [Tibet 92°E]

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<td>-64.4</td>
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<tr>
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<td>18.6</td>
<td>8.8</td>
<td>-6.6</td>
<td>-31.4</td>
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<td>-77.0</td>
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<tr>
<td>Entobbe 0.1°N</td>
<td>18.6</td>
<td>8.2</td>
<td>-7.0</td>
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<td>-54.2</td>
<td>-58.4</td>
<td>-77.6</td>
<td>1960-62</td>
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Tab. 3d Meridional Temperature Cross-Section above Western Africa (along 5°E) and Central Africa (along 16°E)

<table>
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<tr>
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<th>Period</th>
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<tbody>
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<td>14.5</td>
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<td>-46.1</td>
<td>-57.6</td>
<td>-66.9°C</td>
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<td>Tamanrasset 22.0°</td>
<td>28.9</td>
<td>12.8</td>
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<td>-28.8</td>
<td>-47.6</td>
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<td>-70.3</td>
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<tr>
<td>Niamey 13.5°</td>
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<td>9.4</td>
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<td>7.6</td>
<td>-6.7</td>
<td>-31.6</td>
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<td>-72.8</td>
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<tr>
<td>Pt. Lamy 12.1°N</td>
<td>19.7</td>
<td>9.4</td>
<td>-6.5</td>
<td>-30.2</td>
<td>-51.6</td>
<td>-64.6</td>
<td>-73.2</td>
<td>1959-62</td>
</tr>
<tr>
<td>Bangui 4.1°N</td>
<td>17.3</td>
<td>8.2</td>
<td>-6.8</td>
<td>-31.2</td>
<td>-52.0</td>
<td>-65.2</td>
<td>-73.0</td>
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</tr>
<tr>
<td>Leopoldville 4.3°S</td>
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<td>8.9</td>
<td>-6.6</td>
<td>-33.4</td>
<td>-55.6</td>
<td>-67.4</td>
<td>-73.2</td>
<td>1953-58</td>
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</table>
Tab. 4a Zonal Temperature Cross-Section along 25°N, July-August

<table>
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<th>500</th>
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<th>200 mb</th>
<th></th>
</tr>
</thead>
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<td>50.6°E</td>
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<td>15.8</td>
<td>-3.9</td>
<td>-27.2</td>
<td>-48.3 °C</td>
<td>Arabia</td>
</tr>
<tr>
<td>Karachi</td>
<td>24.9°</td>
<td>67.8°</td>
<td>22.2</td>
<td>14.1</td>
<td>-3.5</td>
<td>-25.8</td>
<td>-46.8</td>
<td>Pakistan</td>
</tr>
<tr>
<td>Jodhpur</td>
<td>26.3°</td>
<td>73.0°</td>
<td>24.5</td>
<td>15.0</td>
<td>-0.2</td>
<td>-21.8</td>
<td>-42.5</td>
<td>India</td>
</tr>
<tr>
<td>Allahabad</td>
<td>25.4°</td>
<td>81.7°</td>
<td>23.2</td>
<td>14.1</td>
<td>40.4</td>
<td>-23.2</td>
<td>-43.2</td>
<td>India</td>
</tr>
<tr>
<td>Guwahati</td>
<td>26.2°</td>
<td>91.7°</td>
<td>21.8</td>
<td>13.0</td>
<td>-0.9</td>
<td>-24.0</td>
<td>-42.8</td>
<td>India</td>
</tr>
<tr>
<td>Teungchung</td>
<td>25.1°</td>
<td>98.5°</td>
<td>-</td>
<td>11.0</td>
<td>-2.5</td>
<td>-25.8</td>
<td>-47.0</td>
<td>Chinese P. R.</td>
</tr>
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</table>

Tab. 4b Temperature Distribution along the thermal Equator, July-August

<table>
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<th>Lat.</th>
<th>Long.</th>
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<th>200</th>
<th>100 mb</th>
</tr>
</thead>
<tbody>
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<td>22.8°N</td>
<td>5.5°E</td>
<td>23.9</td>
<td>12.8</td>
<td>-6.6</td>
<td>-28.8</td>
<td>-47.6</td>
</tr>
<tr>
<td>Assan</td>
<td>24.0°</td>
<td>32.0°</td>
<td>24.0</td>
<td>12.3</td>
<td>-3.5</td>
<td>-28.6</td>
<td>-49.4</td>
</tr>
<tr>
<td>Bahrain</td>
<td>26.3°</td>
<td>50.6°</td>
<td>28.9</td>
<td>15.8</td>
<td>-3.9</td>
<td>-27.2</td>
<td>-48.3</td>
</tr>
<tr>
<td>New Delhi</td>
<td>28.6°</td>
<td>77.2°</td>
<td>24.5</td>
<td>14.3</td>
<td>-0.2</td>
<td>-23.4</td>
<td>-44.0</td>
</tr>
<tr>
<td>Hsinchu</td>
<td>27.9°</td>
<td>102.3°</td>
<td>-</td>
<td>12.1</td>
<td>-2.0</td>
<td>-25.0</td>
<td>-46.2</td>
</tr>
<tr>
<td>Pacific 1)</td>
<td>7.8°</td>
<td>-</td>
<td>18.1</td>
<td>9.7</td>
<td>-5.4</td>
<td>-30.8</td>
<td>-53.5</td>
</tr>
<tr>
<td>Tropics 2)</td>
<td>-</td>
<td>-</td>
<td>17.0</td>
<td>8.9</td>
<td>-6.4</td>
<td>-32.5</td>
<td>-54.8</td>
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</table>

1) 6 stations 7 - 10°N, 130 - 170°E, 1957-59
Tab. 5  Temperature and Relative Humidity above Tibet and Adjacent Regions.
July-August 1957-62 (partly incomplete), 00 h GMT.

<table>
<thead>
<tr>
<th>Lat. Long. Height</th>
<th>Average Temperature (°C)</th>
<th>Rel. Humidity [%]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mangya 37.8°N 91.6°E 3000 m</td>
<td>9.9 -4.9 -15.8 -23.0 -45.8 -67.6</td>
<td>44 44 44 53</td>
</tr>
<tr>
<td>Karaz 36.2° 91.6° 2850</td>
<td>12.4 -4.0 -14.2 -26.7 -45.8 -69.2</td>
<td>42 47 45 57</td>
</tr>
<tr>
<td>Yushu 32.1° 96.6° 3873</td>
<td>-2.7 -12.7 -25.8 -47.2 -72.4</td>
<td>80 69 67</td>
</tr>
<tr>
<td>Heihe 32.0° 92.1° 4160</td>
<td>-9.6 -11.2 -23.9 -45.8 -73.6</td>
<td>78 71 59</td>
</tr>
<tr>
<td>Changtu 31.2° 97.3° 3200</td>
<td>-1.4 -11.3 -23.9 -45.8 -71.0</td>
<td>77 72 58</td>
</tr>
<tr>
<td>Lhasa 29.7° 91.0° 3658</td>
<td>-9.7 -10.8 -23.8 -45.6 -73.2</td>
<td>82 74 63</td>
</tr>
<tr>
<td>Haikou 27.9° 102.3° 1599</td>
<td>22.1 -2.0 -11.4 -25.0 -46.2 -72.6</td>
<td>33 71 57 54</td>
</tr>
<tr>
<td>Tongchung 25.1° 98.5° 1628</td>
<td>11.0 -2.4 -11.8 -25.4 -46.7 -74.7</td>
<td>92 78 68 57</td>
</tr>
</tbody>
</table>

Tab. 6  Standard Deviation of Temperature, July-August

<table>
<thead>
<tr>
<th>Lat. (°N)</th>
<th>700</th>
<th>500</th>
<th>400</th>
<th>300</th>
<th>200</th>
<th>100 mb</th>
<th>n(500 mb)</th>
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</thead>
<tbody>
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<td>2.4</td>
<td>3.2°C</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Changtu 31.2</td>
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<td>2.2</td>
<td>2.1</td>
<td>2.6</td>
<td>297</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lhasa 29.7</td>
<td>-1.3</td>
<td>1.7</td>
<td>2.4</td>
<td>2.9</td>
<td>320</td>
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<tr>
<td>Tongchung 25.1</td>
<td>1.3</td>
<td>1.0</td>
<td>1.9</td>
<td>2.0</td>
<td>296</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

For comparison:
| Benina 32.1 | 2.5 | 2.6 | 2.9 | 2.8 | 3.1 | 310 |
| Bahrain 26.8 | 2.2 | 2.6 | 2.3 | 2.8 | 3.6 | 287 |
| Khartoum 15.6 | 1.4 | 1.9 | 2.2 | 3.0 | 171 |
| Aden 12.7 | 1.8 | 1.8 | 1.9 | 2.2 | 3.3 | 202 |
| Lagos 8.6 | 1.2 | 1.4 | 1.2 | 1.7 | 2.8 | 246 |
### Tab. 7 Statistical Wind Parameters over Indo-Pakistan (1958-62)

<table>
<thead>
<tr>
<th>Height [km]</th>
<th>July</th>
<th>August</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>$\sigma_x$</td>
<td>$\sigma_y$</td>
</tr>
<tr>
<td></td>
<td>$\alpha$</td>
<td>$\sigma_x$</td>
</tr>
<tr>
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<td>12.2</td>
<td>114° 7.6</td>
</tr>
<tr>
<td></td>
<td>14.2</td>
<td>101° 12.1</td>
</tr>
<tr>
<td></td>
<td>16.3</td>
<td>76° 22.0</td>
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<tr>
<td></td>
<td>18.4</td>
<td>85° 29.1</td>
</tr>
<tr>
<td>Karachi (1960-62)</td>
<td>12.2</td>
<td>99° 21.2</td>
</tr>
<tr>
<td></td>
<td>14.2</td>
<td>100° 25.3</td>
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<tr>
<td></td>
<td>16.3</td>
<td>101° 30.6</td>
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<tr>
<td>Nagpur</td>
<td>12.2</td>
<td>85° 35.3</td>
</tr>
<tr>
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<td>14.2</td>
<td>86° 49.3</td>
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<td>85° 55.0</td>
</tr>
<tr>
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<td>18.4</td>
<td>85° 53.9</td>
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<td>Vishakhapatnam (1960-62)</td>
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<td>14.2</td>
<td>79° 55.5</td>
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<tr>
<td></td>
<td>16.3</td>
<td>85° 58.5</td>
</tr>
<tr>
<td>Madras</td>
<td>12.2</td>
<td>80° 38.7</td>
</tr>
<tr>
<td></td>
<td>14.2</td>
<td>81° 61.1</td>
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<tr>
<td></td>
<td>16.3</td>
<td>87° 61.2</td>
</tr>
<tr>
<td></td>
<td>18.4</td>
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</tr>
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<td>12.2</td>
<td>79° 46.1</td>
</tr>
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<td>14.2</td>
<td>80° 62.7</td>
</tr>
<tr>
<td></td>
<td>16.3</td>
<td>87° 56.5</td>
</tr>
<tr>
<td></td>
<td>18.4</td>
<td>87° 47.7</td>
</tr>
</tbody>
</table>

- $\alpha$ = direction, $v_x$ = velocity of resultant wind (knots); $\sigma_x$, $\sigma_y$ = standard deviation of zonal (meridional) component (knots); $\sigma_{xy}$ = resultant wind (knots); $\rho_{xy}$ = correlation coefficient between both components; $E_o$ = $\sigma^2\{\sigma^2 + v_x^2\}$ contribution of eddies to the total kinetic energy in percent.
Text of Figures

Fig. 1 Upper Air Stations

Fig. 2 Resultant winds and isotachs at 150 mb, isotach axes at 200 and 100 mb, July-August (data from Tab. 1)

Fig. 3 Meridional cross-section of the mean zonal component $\overline{v}_x$ along 78°E, July-August, 1956-59 and 1961-2 (dashed lines)

Fig. 4 Meridional cross-section of the mean zonal and meridional components above Africa (arranged to Long. 25°E), July-August, 12-18 kms.

Fig. 5 Meridional cross-section of mean temperatures (July-August 1957-62), deviations from a standard atmosphere (equatorial Pacific 7-10°N), along 78°E (data from Tab. 3 b and 4 b). Dashed lines: data probably biased by different types of radiosonde.

Fig. 6 Meridional cross-sections of mean geostrophic zonal wind component along 78°E a) registering balloons up to 1943, b) radiosonde 1957-60.

Fig. 7 Idealized cross-circulation at the entrance and exit region of the Tropical Easterly Jet.

Fig. 8 Model of the (computed) cross-circulation in the exit region above Africa.

Fig. 9 Meridional cross-section of the horizontal divergence (above) and the vertical component of winds at both sides of the Intertropical Convergence Zone above the Atlantic (Flohn 1957)

Fig. 10 Meridional section of the mean annual variation of rainfall along Long. 32°E (Nile valley).
Resultant Winds and Isotachs (kts) 150 mb

Fig. 2
Fig. 4

$\bar{V}_y$ (Africa)
July-August (kts)

$\bar{V}_x$

$5^\circ$N $10^\circ$ $15^\circ$N
Geostrophic Zonal Components
India (75°-90°E), Juli-August

Fig. 6
$J_{ET}$ = Tropical Easterly Jet  \quad J_{WS} = $Subtropical$ Jet (Westerlies)
A) Meridional Circulation (cm s⁻¹)

\[ \text{div}_2 v = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \]
\[ = -0.83 + 2.65 \times 10^{-6} \text{ sec}^{-1} \]
\[ = +1.82 \]

B) Cross-Circulation along 20°E (cm s⁻¹)

Fig. 8
Fig. 9
Annual Trend of Rainfall along 32°E

Fig. 10