LONG PERIOD WIND FLUCTUATIONS IN THE TROPICAL STRATOSPHERE

by

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Submitted to the Department of Meteorology on 22 August 1966 in partial fulfillment of the requirements for the degree of Doctor of Philosophy.

ABSTRACT

The purpose of the thesis is to examine the long period variations in the atmosphere associated with the so-called "biennial oscillation" from an observational point of view, with hopes of gaining insight into the physical mechanisms involved. The variations are not assumed to be periodic; spectral and harmonic analysis techniques are not used in presenting the data.

An examination of daily wind data for the tropical stratosphere shows that the synoptic scale features, which appear to take the form of shear lines, represent truly small perturbations imbedded in a strong, slowly varying, mean zonal flow. Hence the monthly mean zonal wind is a meaningful statistic. The relatively small variation of monthly mean zonal wind with longitude in the same region suggests that zonal averaging is also a meaningful procedure.

Monthly mean, zonally averaged zonal wind data for the region between 32°N and 20°S, 100-7 mb are presented for the general period 7/57-12/64, in the form of time-height sections and meridional cross-sections. The features associated with the "biennial wind oscillation" emerge in considerable detail. Even subtle features appear consistently in independent data from a wide range of latitudes. The wind regimes repeat themselves at two year intervals prior to 1963, but since that time the "period" has lengthened considerably. The downward propagation of the wind regimes is essentially synchronous at all latitudes within 20° of the equator. There is also evidence of a semi-annual variation in the winds at subtropical latitudes.

The momentum balance equation is derived by performing monthly and zonal averaging on the equation of motion for the zonal component. It is shown that the zonal accelerations are determined, for the most part, by the divergence of the horizontal eddy momentum flux and by the mean meridional and vertical motions. Data are presented which show that long period variations in eddy fluxes can account for the appearance of a series of alternating easterly and westerly wind regimes above 20 mb. It is deduced that at the equator, advection by mean vertical motions is responsible for the observed downward propagation. At other latitudes mean meridional motions may also be important.

It is shown that the long period variations in the mean zonal flow are geostrophic to well within 1 km of the equator, and thus wind fluctuations cannot exist without corresponding temperature fluctuations. It is suggested that changes in the zonal wind field give rise to mean vertical motions, which, in turn, produce the temperature fluctuations. Energy considerations indicate that the reverse sequence of cause and effect is extremely unlikely. On this basis it is argued that radiation plays a dissipative role in the "biennial oscillation", rather than a causative one.

The variations in the eddy fluxes responsible for the zonal wind fluctuations in the tropics apparently extend into middle latitudes with large amplitudes, above 50 mb. Since the summer fluxes are negligible, these variations are due to year to year differences in the winter fluxes. There appears to be an overall winter to winter modulation of the intensity of the eddy fluxes and this is reflected in the poleward transports of other quantities, including ozone. Long records of ozone data indicate that there has been a tendency for a two year periodicity in the intensities of the winter seasons, particularly during the decade 1953-63. Possible mechanisms for producing such a periodicity are discussed. It is suggested that interactions between the eddy circulations in middle latitudes and the mean zonal flow in the tropics could possibly produce the observed effects.

Various other manifestations of the "biennial oscillation" in the stratosphere and troposphere are examined. It is shown that a reported two year periodicity in temperature in the polar stratosphere results from year to year differences in the timing of the final warmings. Data on two year periodicities in zonal winds in the middle latitude stratosphere suggest different processes responsible in the two hemispheres. These effects, together with periodicities near two years in certain tropospheric parameters appear to be fundamentally related to the long period variations under investigation.

Interhemispheric differences and relations between troposphere and stratosphere are discussed. The troposphere is considered as a possible source of the long period variations. It is concluded that, although the troposphere is the energy source for the circulations of the lower stratosphere, and hence, in a manner of speaking, for any variations which occur in them, the stratosphere probably holds the key to the cause of the "biennial oscillation".

Thesis Supervisor: Reginald E. Newell

Title: Associate Professor of Meteorology

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It has been my pleasure to do this thesis under the guidance of Professor Reginald E. Newell, who has been both a conscientious advisor and a highly respected colleague in this work. I also appreciate the interest which Professor Victor P. Starr has shown in this study and have profited from the many discussions we have had on the subject. Dr. Robert Dickinson and Messrs. James Miller and John Kidson have also contributed helpful suggestions and criticisms.

I was relieved of many of the data processing problems in having the help of Mr. Charles Nason, who performed much of the data handling, Miss Judith Roxborough and Mrs. Judith Copeland, who wrote most of the necessary computer programs, and Miss Isabel Kole, who drafted the diagrams. Mr. Howard Frazier and the staff of Travelers Research Corporation provided considerable technical assistance in processing the data from the MIT library, and the staff of the MIT Planetary Circulations Project were instrumental in performing many of the computations and in typing the manuscript. In the preparation of this thesis I have relied heavily upon the assistance of all these people.

The Atomic Energy Commission provided financial support which covered the data acquisition and processing costs. These costs were minimized by the generous allowance of time which the MIT Computation Center made available free of charge. During the course of this work my tuition and living expenses were paid for a year by the Ford Foundation and for two years by the Fannie and John Hertz Engineering Scholarship Foundation.

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CHAPTER I

INTRODUCTION

1.1 Discovery of the "Biennial Oscillation"

The explosion of Krakatoa in 1882 produced a dust cloud dense enough to cause optical and climatological effects throughout the world for a number of years. The spread of this cloud during the first few weeks of its lifetime gave scientists their first view of the wind circulations of the tropical stratosphere. "Apart from offshoots toward Japan and South Africa immediately after the explosion, the main body of the cloud moved from east to west at an average speed of 73 miles per hour, completing at least two circuits of the earth in equatorial latitudes". (Wexler (1951)). Von Berson's upper air studies over Africa in 1908-9 with balloons unexpectedly showed evidence of a westerly circulation in the same region. From that time, until the late 1950's the belief in the coexistence of the two opposing flow regimes prevailed. Palmer (1954) described Von Berson's westerlies as "a narrow 'thread' of steady winds whose axis lies at about 2^{O}N and whose base lies near 20 km. The upper transition to the Krakatoa winds varies from month to month and year to year". With the advent of more regular soundings evidence on the variability of the wind structure increased. Korshover (1954) showed that the transition level between westerlies and easterlies varied between 21 and 27 km in a series of observations spanning several years. McCreary (1959) followed the movement of this transition level over the course of several years and found that it moved downward until the upper regime eventually replaced the lower. This led him to suggest that a dynamic view of the tropical stratospheric circulation better fitted the observations than the traditional steady state description. A few months later, Reed (1960) announced the discovery of what was to be called the biennial or 26 month oscillation in the wind structure of the tropical stratosphere.

1.2 Comments on Previous Research

Since the time of their discovery it has usually been assumed, explicitly or implicitly, that these long period wind fluctuations are periodic, or at least quasi periodic, and the use of the language of the harmonic oscillator has become common in both observational and theoretical works on the subject. Harmonic and spectral analysis techniques have become the tools of the trade, so to speak, with amplitude, phase and period information taking precedence over untreated time series as a form of data presentation.

The use of harmonic and spectral analysis techniques has been subject to shortcomings in some cases. There is no doubt that harmonic analysis, when used correctly, can be a useful tool for obtaining amplitude and phase information on cycles known to be present in a data record. However, it must be used with caution in cases where periodic behavior is only suspected. Its very use involves the assumption that the periodicity actually exists, and it does yield amplitude and phase information, regardless of whether the assumption is correct. Power spectrum analysis is a more objective means of obtaining information on periodic elements in a data record, but its application to short records will not yield reproducible results unless the cycle in question is truly periodic. In addition to these shortcomings both types of analysis suffer from the fact that they are so highly specialized, being designed, as they are, to yield information specifically related to the periodic elements present in time records. In singling out this kind of information, they ignore all other details inherent in the data. There are times when the resulting simplification is highly desirable. However, in data which contain much information which has yet to be assimilated, analysis by these techniques alone may leave important features unnoticed.

1.3 Purpose of this Thesis

Many of the papers on this subject which have employed harmonic and/or power spectrum analysis have done so judiciously and have obtained information which is reliable and useful. However, it is the contention of the author that the traditional view of these long period variations as a periodic phenomenon, whose period varies from one cycle to the next leaves something to be desired, and therefore it may be profitable to look at the data without any preconceived notions regarding periodicities. The purpose of this thesis is to provide an unsophisticated presentation of the data relevant to the subject with a view toward an understanding of the basic physical processes involved. Consequently, the use of the language of harmonic oscillators is avoided, except where it is possible to use it in the precise sense (e.g., in relation to the annual cycle), or where its usage is necessary to refer to material in the literature. The data are presented in the simplest possible form. Twelve month running means is the most complicated type of smoothing used, and even that is avoided wherever possible.

1.4 Preview of the Subject Material

Chapter II provides some background material on the day to day variations of the wind in the tropical stratosphere. This brief excursion from the main theme of the thesis is justified by the need to investigate (1) to what extent monthly means may be taken to represent the situation on individual days within the month and (2) the type of disturbances which are responsible for the eddy transports of momentum in this region. The treatment is intended only as a pilot study of the daily wind data; a more detailed study would have to deal with some serious problems from the standpoint of data availability.

Having obtained information on these items, the discussion returns to the main theme of the thesis. The long period wind variations of the tropical stratosphere are considered in Chapter III. After a detailed presentation of the mean monthly zonal wind data of the region, the variations are examined from the standpoint of the requirements of the momentum budget, geostrophic balance and the heat budget, respectively. It is found that the simultaneous satisfaction of these requirements can be obtained, given a specified field of mean meridional motions and appropriate values of the relevant radiative parameters. Energy considerations suggest that long period variations in the eddy fluxes of momentum are the immediate cause of the observed variations of wind and temperature in the tropics.

The quest for a cause of the long period variations in the momentum fluxes leads to an investigation of the seasonal circulations of the stratosphere at middle latitudes in Chapter IV. It is found that at least in the northern hemisphere, the winter seasons exhibit year to year differences in the intensity of eddy activity. The nature, and possible causes of this phenomenon are discussed and it is concluded that the atmosphere is capable of producing the observed effects through interactions between the mean and eddy circulations.

Chapter V contains a survey of various other parameters outside the tropical stratosphere which show signs of being connected with the so called 26 month oscillation. The multiplicity of these parameters precludes any possibility of an all-embracing mechanism which could be called upon to account for all of them. However, wherever possible suggestions are made as to possible linkages between the more closely related phenomena.

The concluding chapter (VI) discusses certain questions concerning the global nature of this phenomenon, such as the relationship between northern and southern hemispheres and between troposphere and stratosphere, and the ultimate cause of long period variations in the atmosphere.

CHAPTER II

SYNOPTIC SCALE DISTURBANCES IN THE TROPICS

Although the tropical stratosphere has been the subject of considerable attention with regard to long period wind fluctuations, it has been almost entirely neglected as far as synoptic scale disturbances are concerned. This neglect is probably not so much due to oversight as simple lack of interest. The weak, barely noticeable disturbances of this region are far less intriguing than the distinctive and sometimes spectacular features of the polar vortices at higher latitudes. And to a student of the tropics, the variety of phenomena in the lower troposphere which exert a far more direct influence on man's environment provides a virtually limitless field for research. Even if there were more reason to be interested in the tropical stratosphere lack of data presents an almost insurmountable obstacle to any detailed synoptic study. From Fig. A, which shows the location of all the northern hemisphere stations for which data was available for this study it is clear that with the exception of the West Indies and western Pacific, the tropics are almost entirely devoid of stratospheric data. Moreover, the disturbances under consideration are so weak that observational inaccuracies and mesoscale phenomena create a noise level high enough to distort and mask many of

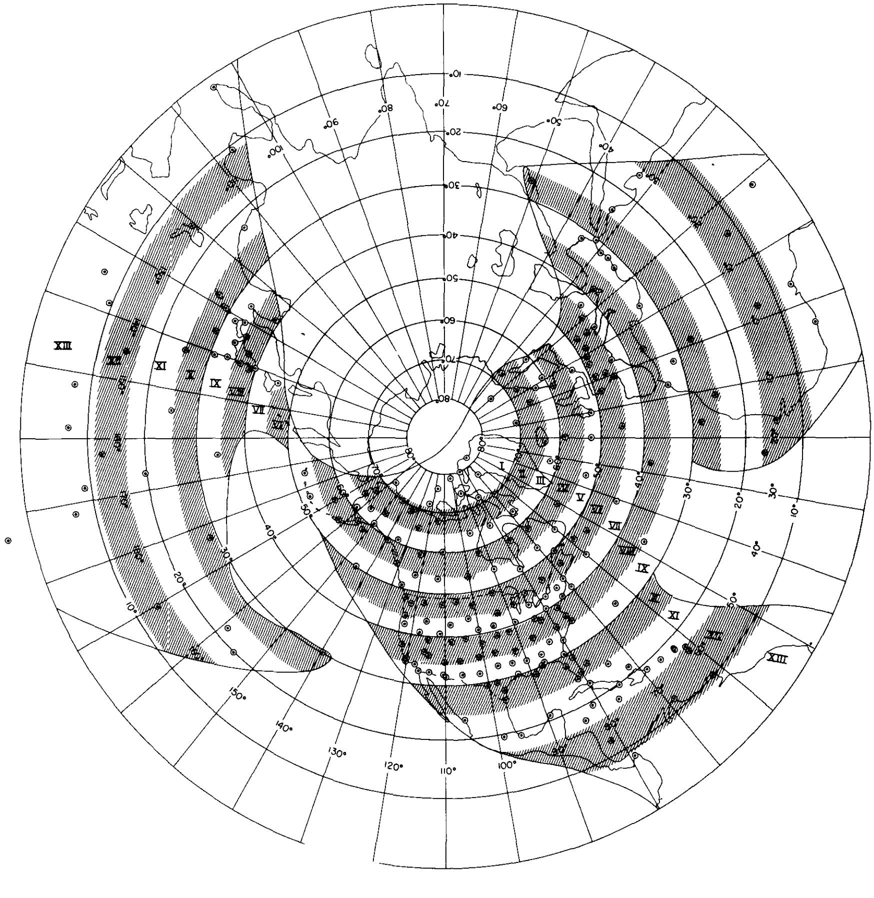


Figure A The distribution of northern hemisphere stations used in this study. The latitude bands into which the stations were grouped are denoted by Roman numerals. Blank areas denote regions with no data.

the synoptic scale features, an effect which is particularly troublesome in the absence of a dense station network. Unless the satellite program makes a real contribution to the acquisition of stratospheric data there is not likely to be any significant improvement in this situation within the forseeable future.

For reasons stated in the introduction this chapter will concern itself with these synoptic scale perturbations. Because of the very major difficulties outlined above, and since this chapter is, in a sense, only background material for the main part of the work to follow, only a rather brief treatment of the subject will be given here. The scarcity of papers on this topic in the literature requires that the discussion be based mainly upon data presented herein.

2.1 Presentation of Daily Wind Data

Time height sections in the 125-10 mb layer were prepared for the zonal and meridional wind components computed directly from daily and sometimes twice daily soundings. In all cases the meridional sections fail to show any coherent structure that is consistent from station to station, which suggests that the noise level is too high to permit an effective representation by this method. The zonal component, on the other hand, shows distinct large scale features which stand out above the noise level.

Fig. 2.1 shows zonal wind sections based on twice daily soundings at Balboa and San Juan. Despite the rather wide geographical separation of the stations, the same features are evident in both sections. The large space scale of these disturbances and the fairly long time scale suggests that a judicious choice of averaging procedures can be applied to the data to reduce the noise level without smoothing out the features under consideration. Accordingly, the data for all the subsequent sections appearing in this chapter were processed in the following manner:

(a) Whenever two or more observations were available at a station on the same day they were averaged together to form a mean for that day

(b) Daily means from the stations indicated on the sections were averaged together to form daily means for the station group.

In order to study the synoptic scale fluctuations in the zonal wind as a function of latitude and season, stations in two regions were examined during one winter month (January, 1964) and one summer month (July, 1964). The station groups are located by necessity, in the two regions of the best data coverage: The West Indies and the western Pacific. In both areas enough data was available to divide the stations into two subgroups; one centered around 10°N and the other near 20°N.

Figure 2.2 contains the winter section for the West Indies region.

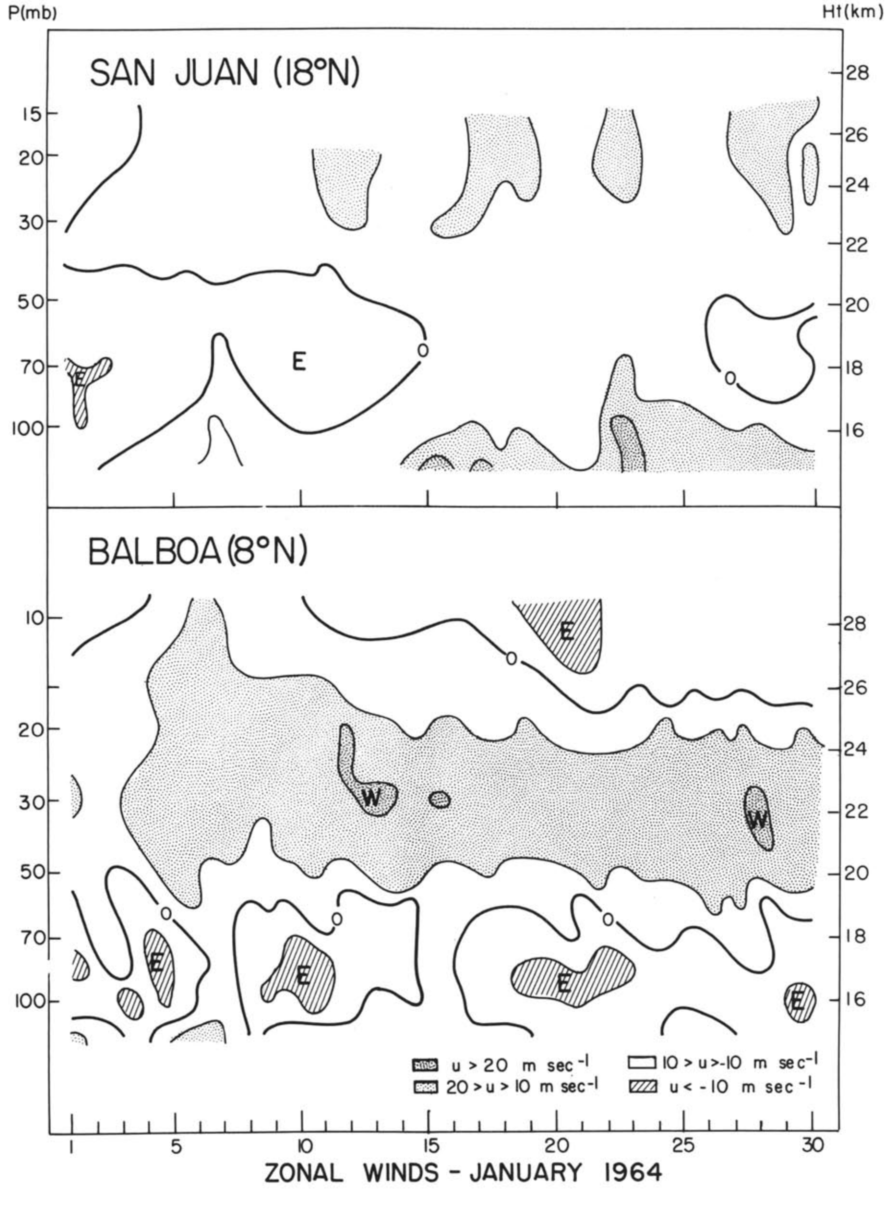


Figure 2.1 Time-height sections of zonal wind for San Juan and Balboa, January, 1964, analyzed from twice daily soundings. Lines are at intervals of 10 m sec^{-1}.

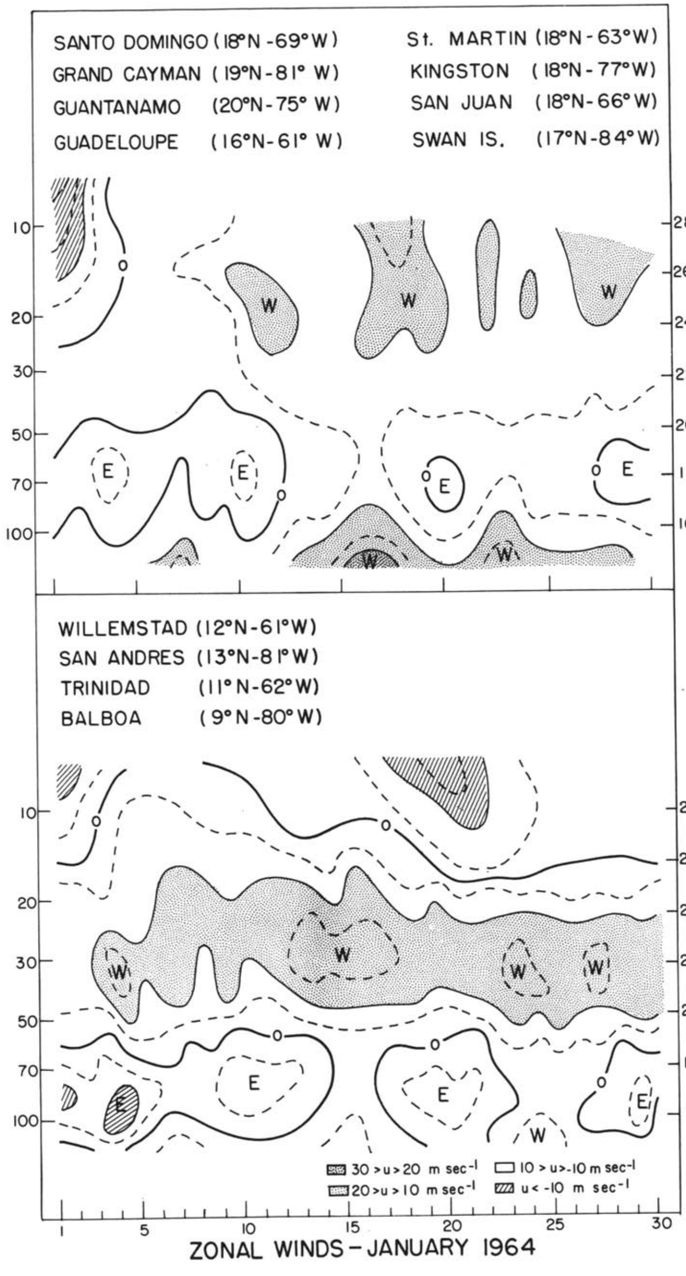


Figure 2.2 Time-height sections of zonal wind for two station groups in the West Indies, January, 1964. Solid lines are at intervals of 10 m sec^{-1}.

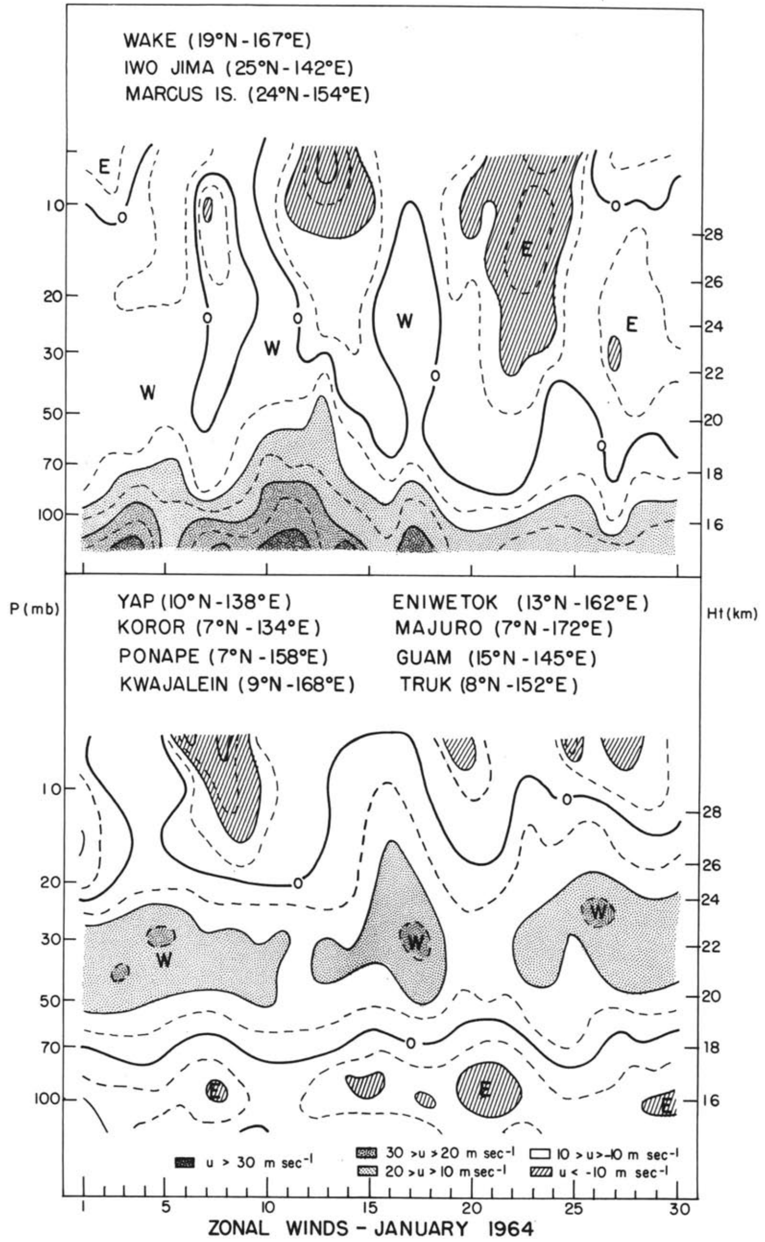


Figure 2.3 Time-height sections of zonal wind for two station groups in the western Pacific, January, 1964. Solid lines are at intervals of 10 m sec^{-1}.

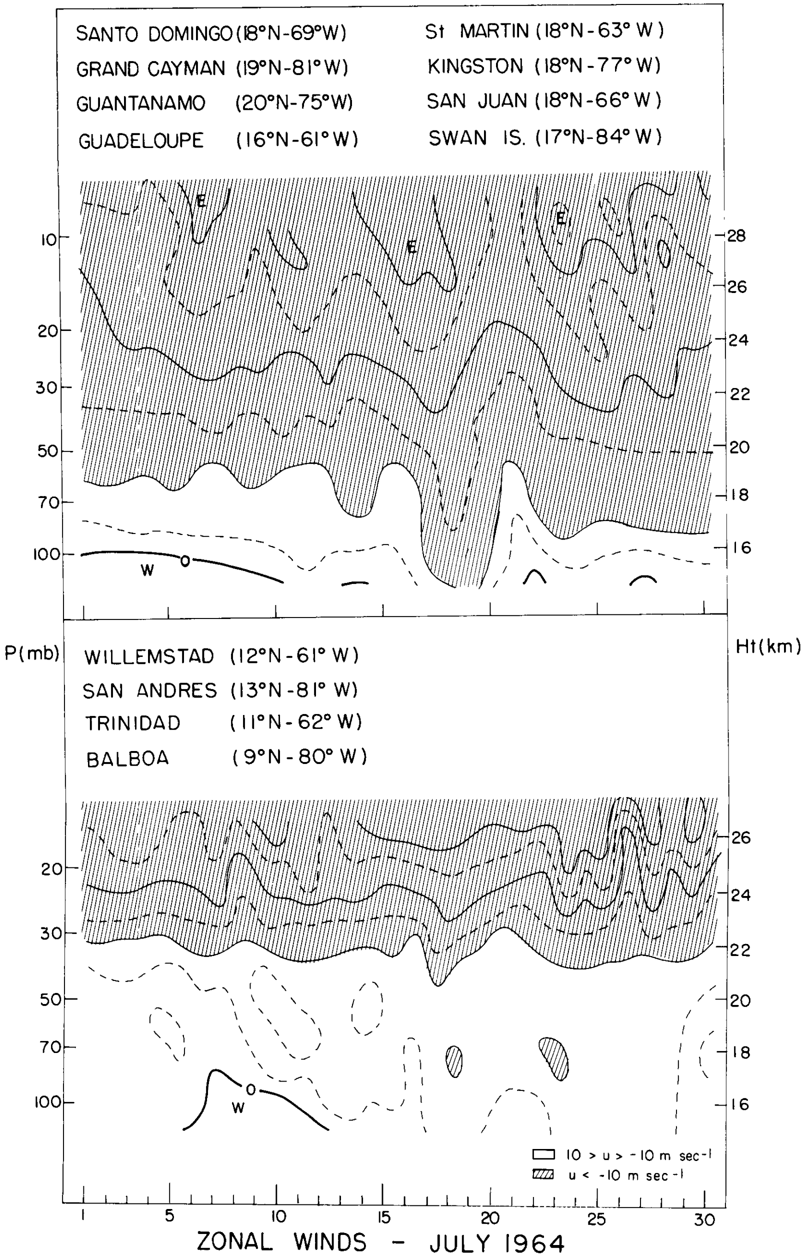


Figure 2.4 Time-height sections of zonal wind for two station groups in the West Indies, July, 1964. Solid lines are at intervals of 10 m sec^{-1}.

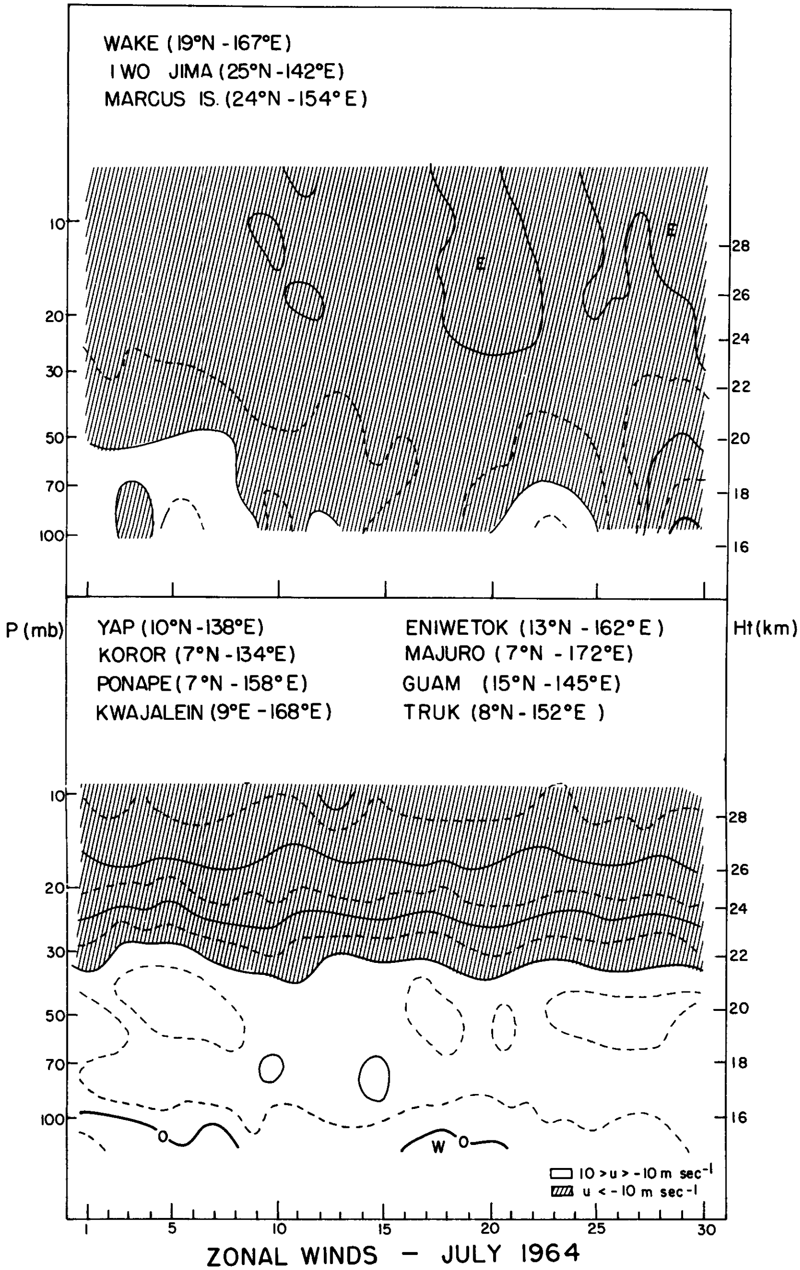


Figure 2.5 Time-height sections of zonal wind for two station groups in the western Pacific, July, 1964. Solid lines are at intervals of 10 m sec^{-1}

It is reassuring to find largely the same features as appear on the individual sections for Balboa and San Juan in Fig. 2.1, but with better resolution. The most marked difference between the sections at different latitudes is in the amount of vertical shear. The western Pacific winte sections in Fig. 2.3 bear the same relationship to one another, the shear also being larger at the lower latitude. The summer sections (Figs. 2.4 and 2.5) differ from the winter sections in that the disturbances are weaker and more sporadic. Aside from these differences the synoptic patterns in all eight sections are quite similar, displaying the common properties:

1. A time scale in the order of one week

2. A horizontal extent of at least 10^{o} in latitude, 30^{o} in longitude

3. A variable range in the vertical. Some disturbances appear to extend from the upper troposphere to the 10 mb level while others are limited to the upper or lower parts of the layer.

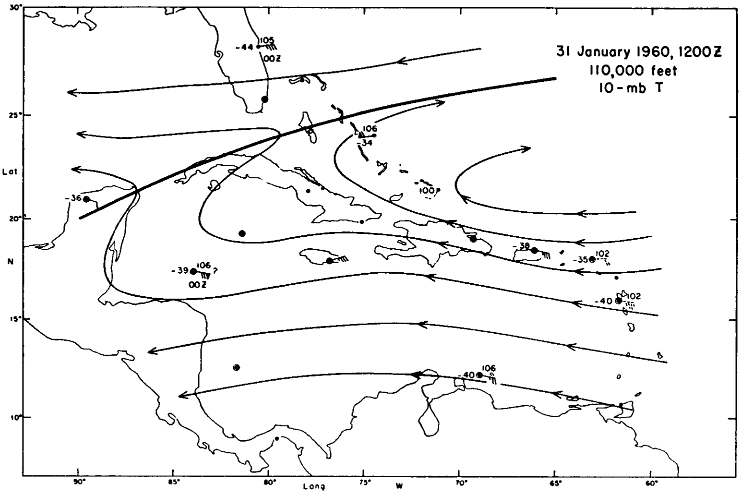
4. Little, if any, phase dependence upon latitude or height. (The resolution of the sections is not fine enough to preclude the possibility of a phase lag or a day or two between the lower and higher latitudes).

5. Larger variations of the zonal wind component than of the meridional wind component.

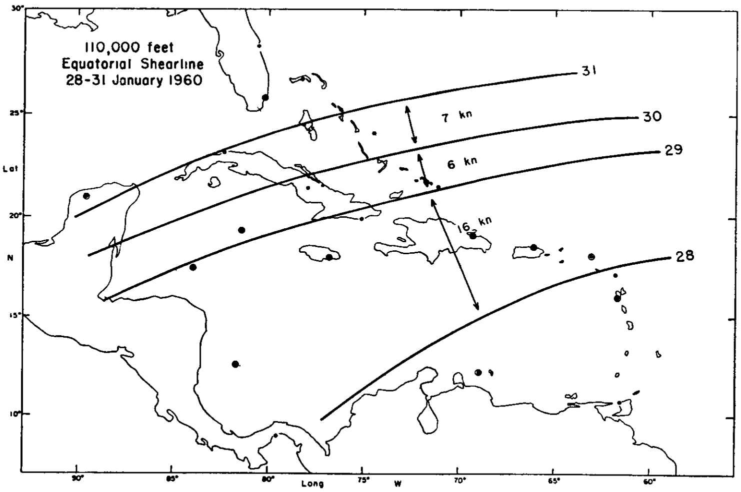
The smallness of the meridional component imposes some limitations on the form which the disturbances may take. It would seem that the wavelike disturbances observed at middle latitudes would be ruled out by such restrictions. The absence of strong meridional shears in the mean winds (see Fig. 3. 4) precludes the possibility that small meanders in a narrow jet could cause variations in the zonal wind without disturbing the meridional wind field. A more likely, and perhaps the only possible form that the disturbances could take would be that of a shear line, similar, in some respects, to those observed in the tropical troposphere. Riehl and Higgs (1960) found that such a model best described an individual disturbance over the West Indies which they analyzed in some detail. Fig. 2. 6 shows a sample of their map analyses. Such disturbances would undoubtedly be capable of transporting momentum, the direction of the transport being determined by the orientation of the shear line.

2. 2 Monthly Statistics

Another measure of the short term variability of the wind is the monthly statistics, particularly the temporal standard deviations of the zonal and meridional wind components. These statistics have been computed, and a detailed presentation of the results is in preparation. Some of the data on the meridional wind component is already available in published form (Newell, Wallace and Mahoney (1966)). A brief summary of the data for both components is given in Table 2. 1 which shows the



Winds at 110,000 ft and 10-mb temperatures, 31 January 1960.



Successive positions of equatorial shearlines 28 to 31 January 1960 and rates of mean 24-hr displacement (knots).

Figure 2. 6. (After Riehl and Higgs (1960)).

Table 2.1. Temporal standard deviations of the wind components and temperature as a function of latitude, height and season.

	σu (m sec^{-1})		σv (m sec^{-1})		σT (^{o}C)	
	Dec-Jan	Jun-Jul	Dec-Jan	Jun-Jul	Dec-Jan	Jun-Jul
8^{o}N						
100 mb	7.5	6.4	5.3	4.5	2.1	2.7
50	4.7	5.7	2.9	3.2	2.1	2.1
30	5.3	5.0	3.3	3.4	2.0	2.0
20		5.7		3.1		1.8
14^{o}N						
100 mb	7.4	6.1	5.4	4.3	1.9	2.2
50	3.8	3.9	2.7	2.8	2.0	1.9
30	4.6	4.0	2.9	2.8	2.0	1.9
20	5.3	4.2	3.2	3.0	2.1	1.9
15	6.2	4.6	3.4	3.0	2.3	2.0
22^{o}N						
100 mb	7.5	5.5	6.0	3.7	2.0	2.0
50	4.2	3.6	3.0	2.6	2.0	1.7
30	4.5	4.0	3.2	2.8	1.9	1.7
20	5.2	4.5	3.2	2.8	2.1	1.7
15	6.2	4.3	3.6	3.2	2.2	1.8
28^{o}N						
100 mb	7.8	6.2	7.0	4.4	2.6	2.2
50	6.3	3.5	3.4	2.4	2.4	1.6
30	6.8	3.9	3.2	2.7	2.1	1.6
20	7.3	4.3	3.7	2.8	2.3	1.9
15	8.1	4.6	4.1	3.1	2.7	2.0

seasonal means for a four year period (5/58-4/62) as a function of latitude, height and season. The following is a brief review of these in light of the model deduced above from a limited inspection of daily data.

1. The standard deviation of the zonal wind component is larger than that of the meridional component at low latitudes by about a ratio of 3:2. This is perhaps not quite as large as would be expected from the time height sections. It is possible that the instrumental and mesoscale* noise level places a lower limit on the values of standard deviations.

2. The standard deviations of both components are largest during the winter seasons; particularly at higher latitudes. During the winter they increase with latitude, while during the summer they do the opposite.

3. At 50 mb and above, typical values for the standard deviation of the zonal component are in the range of 4-5 m sec^{-1}; for the meridional component 3 m sec^{-1}. The former figure agrees with the time-height sections.

4. Variability decreases from 100 to 50 mb, and increases with height above 50 mb, as in the sections.

*Some irregularities in the wind soundings at these levels appear to be real features. For a discussion of these phenomena, see Newell, Mahoney and Lenhard (1966).

CHAPTER III

THE WIND AND TEMPERATURE VARIATIONS IN THE TROPICAL STRATOSPHERE

As far as this study is concerned, the most important result of the examination of daily data carried out in the previous chapter is that the monthly mean wind profile is a close approximation to the instantanious wind profile on a given day, and conversely, that the average of the wind profiles for several scattered days within a month yields an accurate estimate of the monthly mean. In other words, the monthly mean zonal wind is a meaningful and stable statistic for the region of the atmosphere under consideration. This fact is the basis for the extensive use of monthly mean wind data in this chapter.

3.1 The Question of Zonal Averaging

The same criteria may be applied to the process of zonal averaging to determine whether it will prove useful in further reduction of the data. Belmont and Dartt (1964) have analyzed time-height sections for a number of tropical and subtropical stations. Their results show a strong similarity between sections for stations at neighboring latitudes, regardless of longitude. This suggests that despite the very poor longitudinal distribution of stations with available data, it may be possible to derive meaningful zonal averages. This is equivalent to saying that the eddy disturbances appearing on mean monthly maps are small compared to the mean zonal flow.

On this basis, zonal averaging was applied to the data with hopes that the appearance of the resulting sections would further serve to justify the validity of this procedure. Zonal means were formed by apportioning the stations into latitude belts. (see the Appendix for a list of stations in each belt). In the averaging, stations were weighted by the number of observations which they comprise. This eliminates the possibility of giving undue weight to a station with very few observations. The number of observations at individual stations varies considerably from month to month and from level to level. Moreover, not all stations report at the same levels, or from the same period of record. Hence the values at different points in the sections are based on different station distributions. In view of this heterogeneity of the data, the smoothness of the sections in Fig. 3. 1 and 3. 2 is perhaps the best evidence of the validity of the assumption of zonal symmetry. Only poleward of 20° and below 80 mb was any smoothing required in piecing together data from different longitudes.

3. 2 Time-height Sections

Time-height sections as a method of presentation of zonal wind

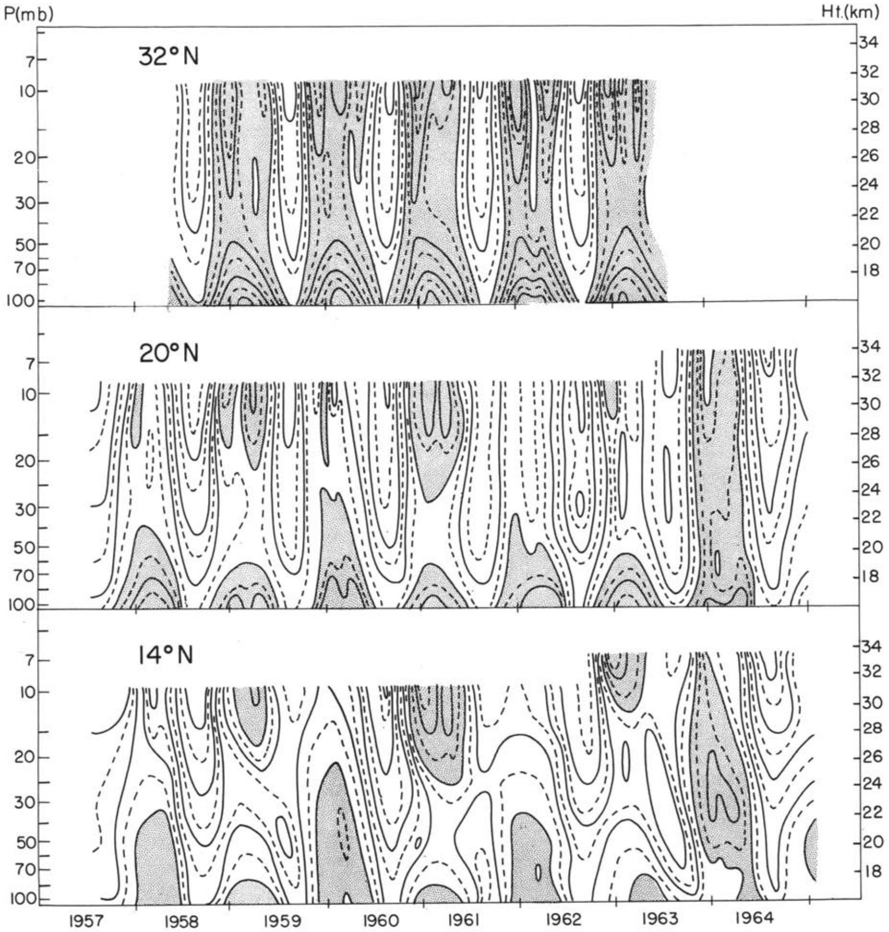


Figure 3.1(a) Time-height sections of zonal wind averaged around latitude circles as indicated. Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies.

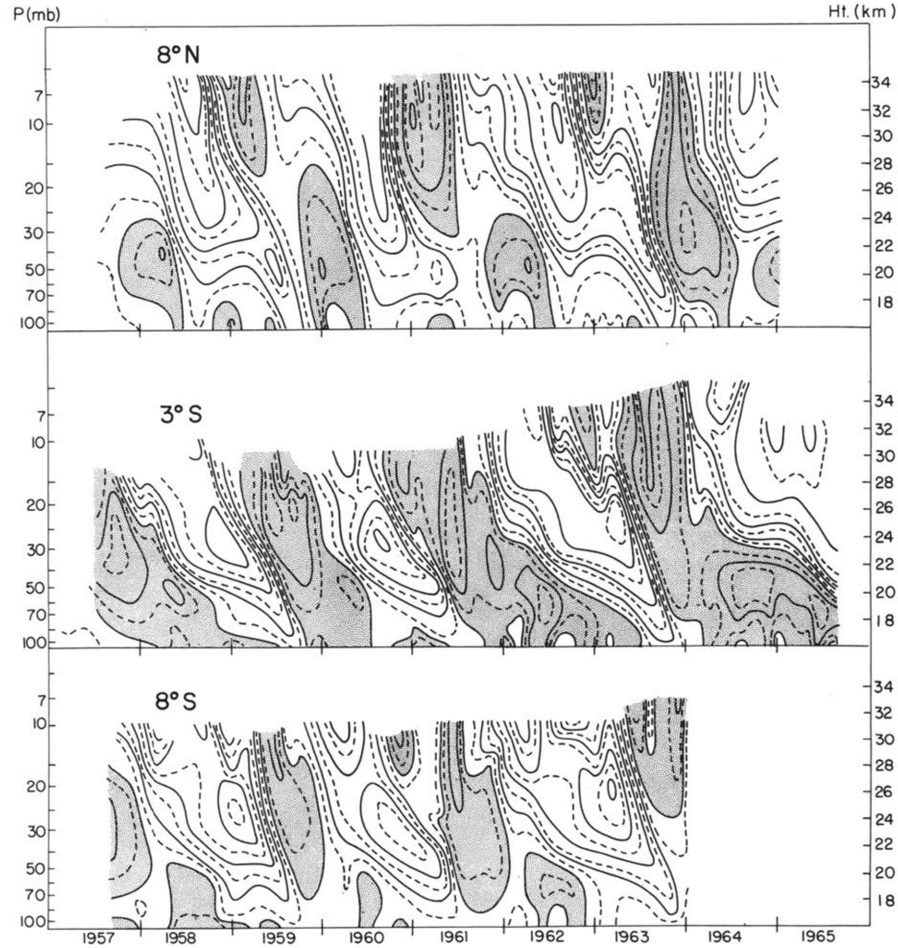


Figure 3.1(b) Time-height sections of zonal wind averaged around latitude circles as indicated. Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies.

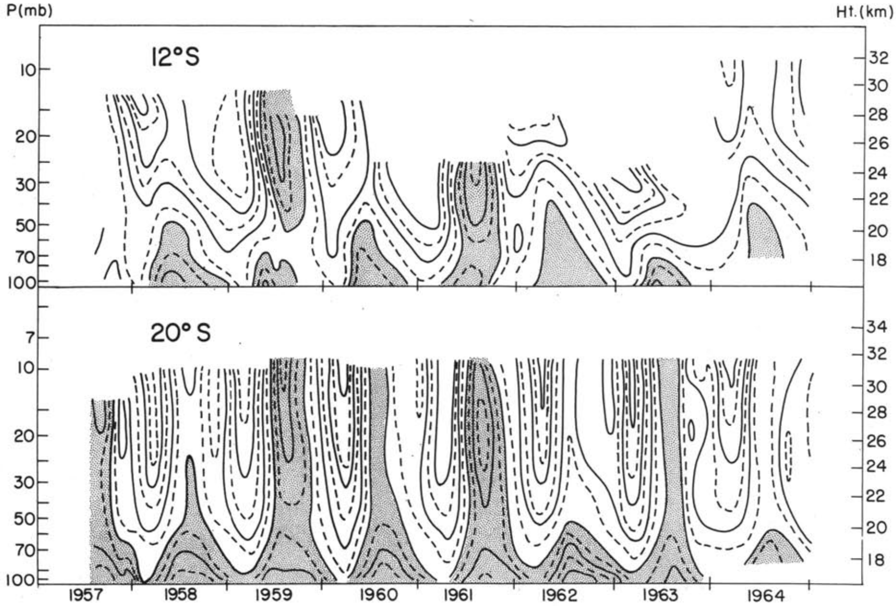


Figure 3.1(c) Time-height sections of zonal wind averaged around latitude circles as indicated. Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies.

data have been employed previously by Reed (1961) and Belmont and Dartt (1964) for individual stations. The justification for their use in the present work is to take advantage of the more extended coverage and improved resolution which the zonally averaged data affords, and to provide a more detailed account of the variation of the wind structure with latitude. A comprehensive view of the behavior of the zonal wind as a function of latitude, height and time will serve as an historical account of the winds in this region for an extended time period and as a convenient starting point for the discussion of the long period fluctuations with which this thesis is primarily concerned. The sections span the period from the beginning of the IGY to the most recent time for which data is available.

The gross features of the biennial oscillation, such as its amplitude and phase distribution and its combination with the annual cycle to produce the effects observed in subtropical latitudes, have been well documented in previous works, (Reed, (1964a), Belmont and Dartt (1964) and no attempt is made to reiterate them here. Emphasis is placed upon some of the more subtle features which are evident in the time-height sections presented herein.

1. The superposition of annual and biennial cycles presents only an approximate description of the zonal wind behavior in this region. Each

individual regime of easterlies or westerlies appears to have its own peculiar characteristics, which are not accidental features of the data sample, but real physical occurrences. For example, the strong easterly regime of 1962-63 and its rapid disappearance during late 1963 are evident in all the sections within 20^{O} of the equator. Even short period features cover a wide range of latitude. For example, the double maximum in the westerlies during the 1960-61 northern hemisphere winter is discernible as far south as 8^{O}N and the strong westerlies which occurred prior to the January 1963 warming and their subsequent disappearance are features which actually appear to extend across the equator at high levels.

2. One feature common to all wind regimes is the tendency for westerlies* to propagate downward more rapidly than easterlies. This is best seen close to the equator where the annual cycle is not present. This difference in propagation rate causes easterly regimes to become progressively flattened in the vertical as the move downward.

3. At very low latitudes the wind regimes prior to the beginning of 1963 repeat themselves at intervals of very close to two years. Since 1963, the duration of the regimes has been decidedly longer; so much so that by 1965 there are easterlies at levels where westerlies would be expected if the two year periodicity had continued.

The annual cycle is clearly the dominant feature in all sections

* the lower or leading edge of a westerly regime

more than 10° of latitude from the equator. Its presence obscures, to a large extent, the other long period features, particularly at subtropical latitudes. Reed (1964a) and Belmont and Dartt (1964) have both attempted to remove the annual component from the data; Reed by a harmonic analysis of the annual and "26 month" cycles, and Belmont and Dartt by subtracting what they estimated to be the annual cycle from the time-height sections for several stations. Both studies showed the amplitude of the long period features to be decreasing with latitude so as to be barely detectable at 20°, and the time sequence of these features, (i. e., "The phase of the biennial or 26 month component") to be basically independent of latitude.

The data available in the present study provides an opportunity to remove the annual component by another method: by averaging pairs of sections at corresponding latitudes in the northern and southern hemispheres. Figure 3. 2 shows the results of applying this procedure to the sections at 8° and 20°, together with the section for 3°S which represents the long period variations near the equator.

The similarity between the 8° section and the equatorial section, even in some of the more subtle details is quite remarkable. That the two sections with completely independent data should show essentially the same features is further proof of the broad extent of the wind variations under consideration and the ability of the data to represent them accurately and in considerable detail with this form of analysis.

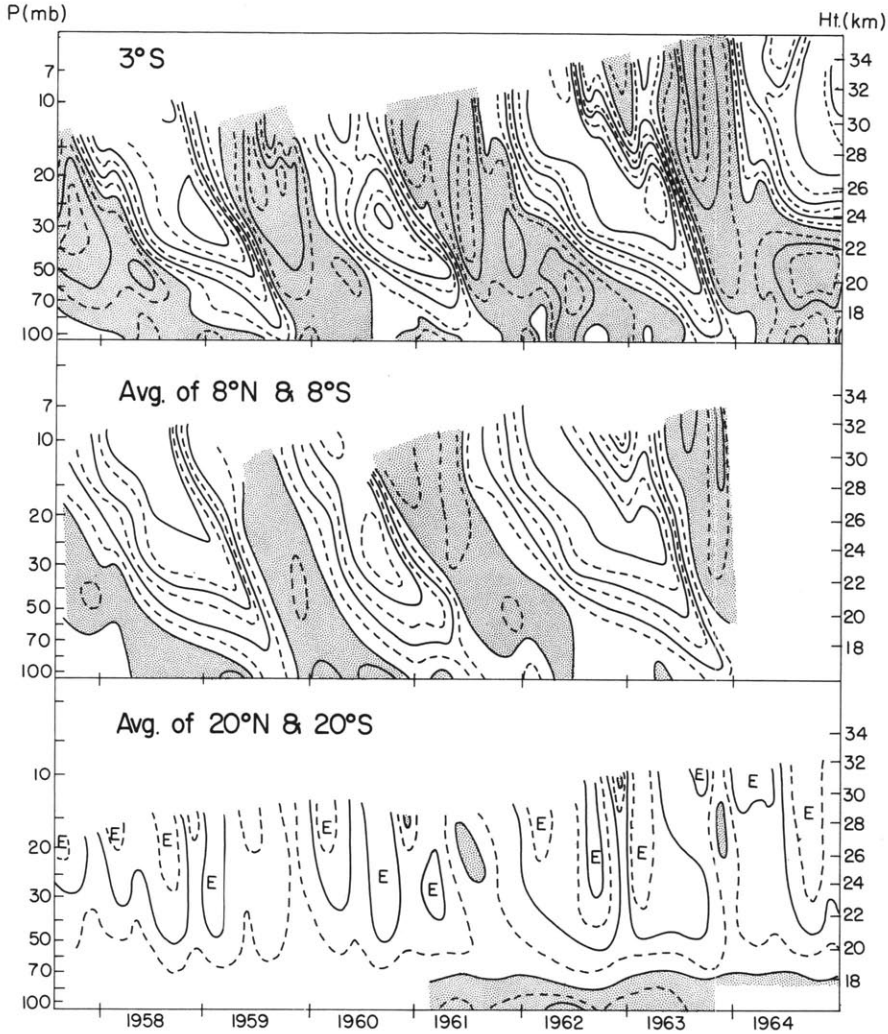


Figure 3.2 Time-height sections of zonal wind averaged around latitude circles as indicated. The lower two sections are the averaged of pairs of sections in Figure 3.1. Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies. December values are placed on tie marks.

The major differences between the equatorial and 8^{o} sections are (a) the decreased amplitude of the features in the 8^{o} section which confirms the results of Reed (1964a) and Belmont and Dartt (1964) and (b) a slightly more westerly long term mean in the equatorial section, an effect which has also been previously noted (Reed (1964a)).

The section for 20^{o} is more irregular, but it is still evidently quite strongly related to the lower latitude sections. In this respect it also confirms the findings of the earlier studies. An unexpected feature noticeable in this section is the suggestion of a semi annual periodicity with easterly maxima in February and August. In retrospect, the lower latitude sections also show some evidence of the same cycle at the highest levels, particularly toward the end of the record where 7 and 10 mb data are more abundant.

The appearance of a semi-annual oscillation in winds above 10 mb in the tropical stratosphere is not without precedent; Reed (1965) suggested the existence of such a cycle on the basis of the rocket data from Ascension Island shown in Fig. 3.3. In this figure the cycle is detectable at the 10 mb level (32 km) and its amplitude increases with height above that level. The phase at 32 km matches well with that at the 10 mb level in Fig. 3.2. Thus the results of the present study may be interpreted as indicating an extension of Reed's semi-annual cycle to lower levels in the subtropics.

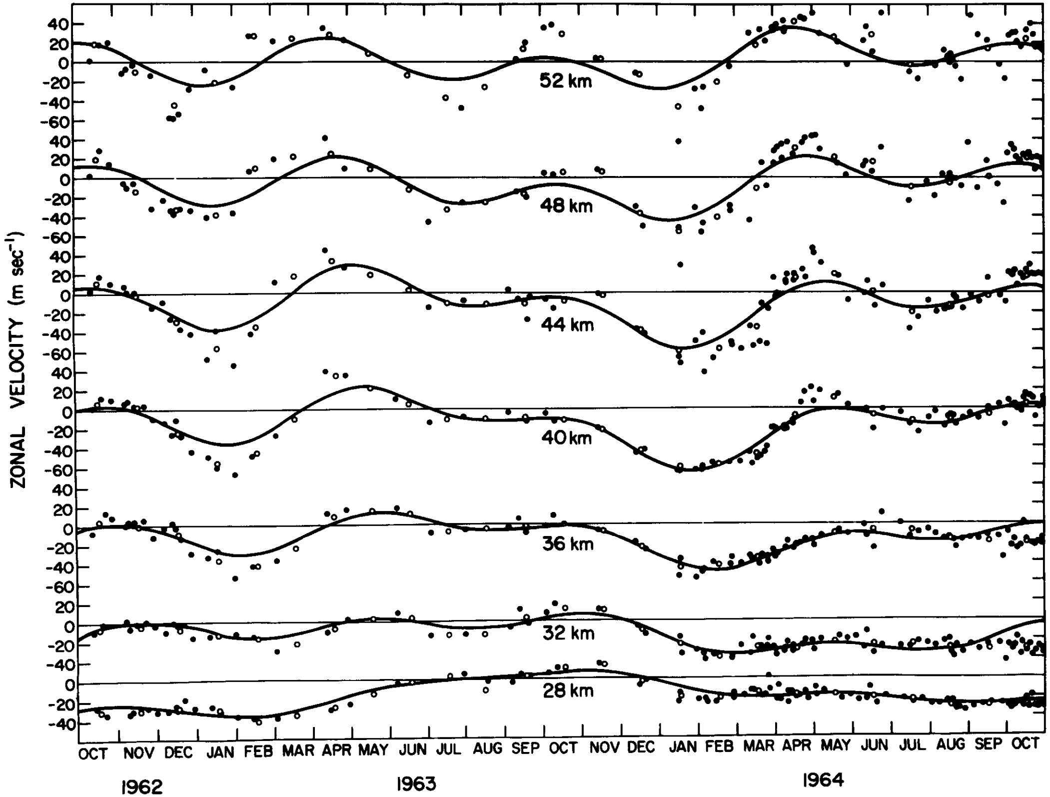


Figure 3. 3 Zonal wind velocity at Ascension Island. Solid circles: individual observations; open circles: monthly means. Curves were drawn objectively by summing the first, second and fourth harmonics obtained from harmonic analyses of the monthly means. (After Reed (1965))

Although it is tempting to speculate on the nature and cause of this semi-annual cycle, it will be difficult to do justice to the subject until rocket data provide a much more comprehensive view of its behavior at high levels in both hemispheres over an extended time period. The present study will concern itself primarily with the longer period features.

3.3 Meridional Cross Sections

In order to provide an instantaneous picture of the wind structure at regular intervals, meridional cross sections have been prepared using the same data as in the time-height sections for alternate months from July 1957 to November, 1963. These are shown in Fig. 3.4.

This sequence supports the contention that the long period wind variations repeat themselves at intervals of two year prior to 1963. The similarity between sections two years apart during that period is quite strongly evident. It is possible to trace the life history of successive regimes in the sections as they propagate downward, which shows that they maintain their identity despite the seasonal reversals associated with the annual cycle.

The slow, systematic evolution of the zonal wind structure of the tropical stratosphere as seen in the time-height sections and the meridional

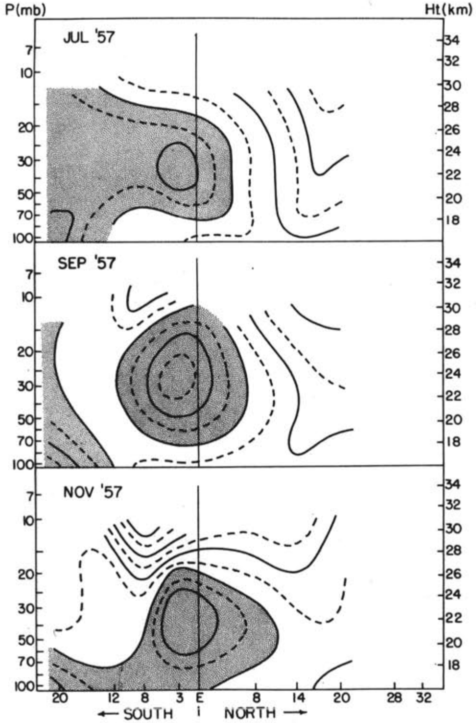


Figure 3.4(a) Meridional cross sections of zonal wind averaged around latitude circles for months indicated. Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies.

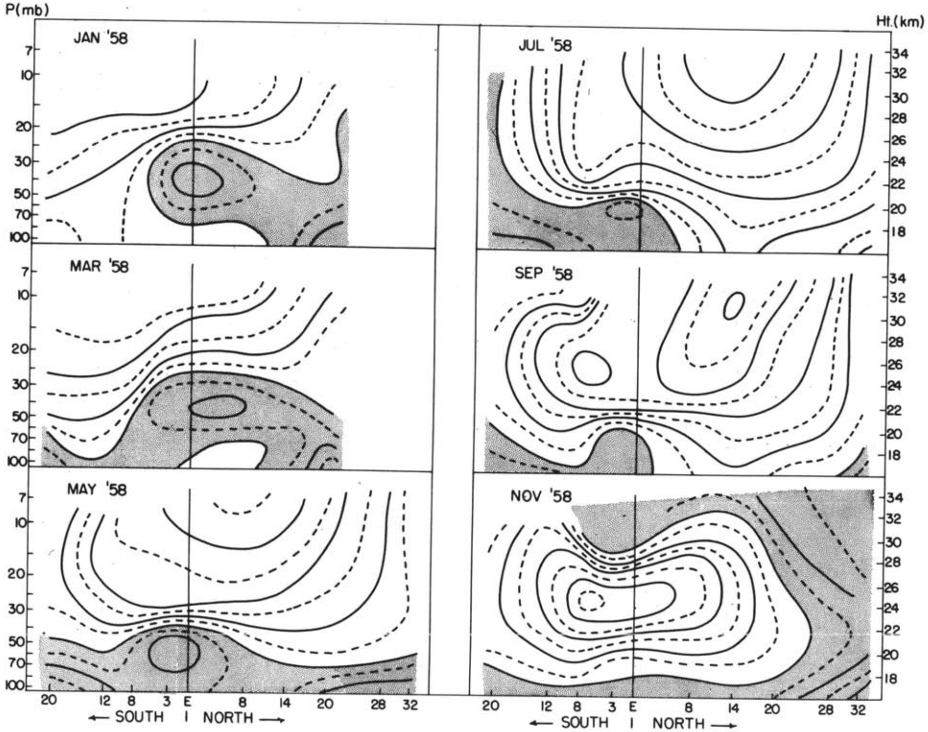


Figure 3.4(b) Meridional cross sections of zonal wind averaged around latitude circles for months indicated. Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies.

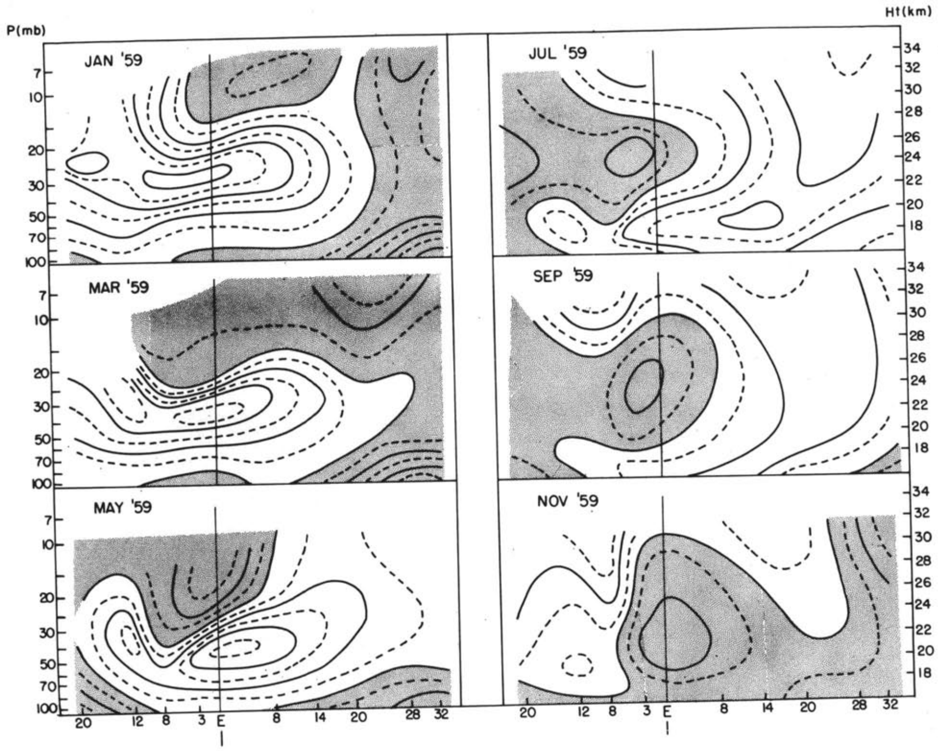


Figure 3.4(c) Meridional cross sections of zonal wind averaged around latitude circles for months indicated. Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies.

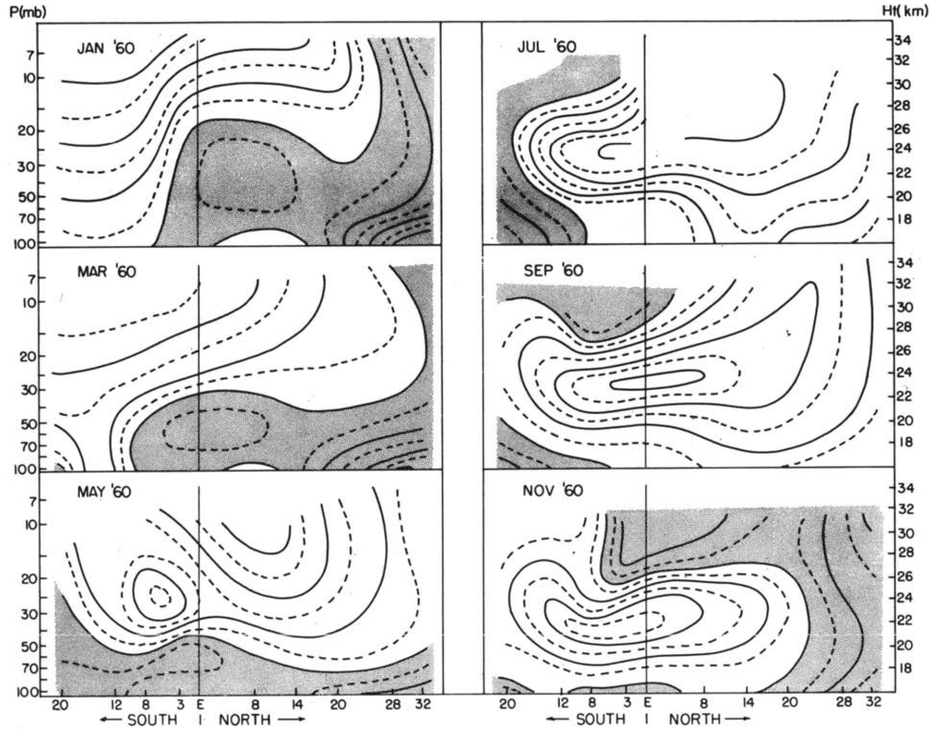


Figure 3.4(d) Meridional cross sections of zonal wind averaged around latitude circles for months indicated. Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies.

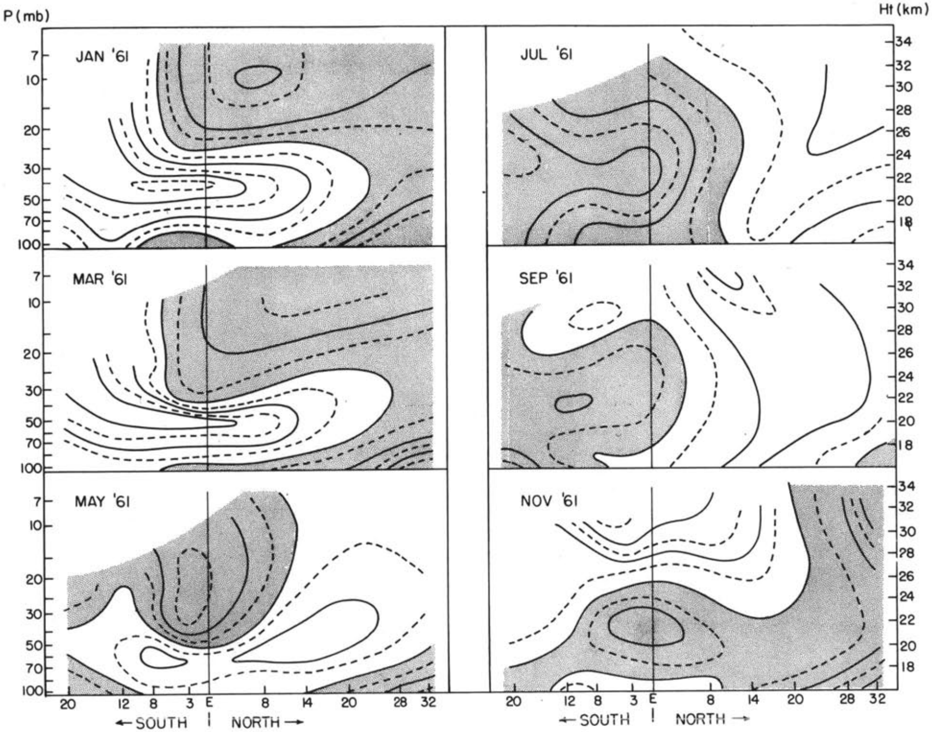


Figure 3.4(e) Meridional cross sections of zonal wind averaged around latitude circles for months indicated. Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies.

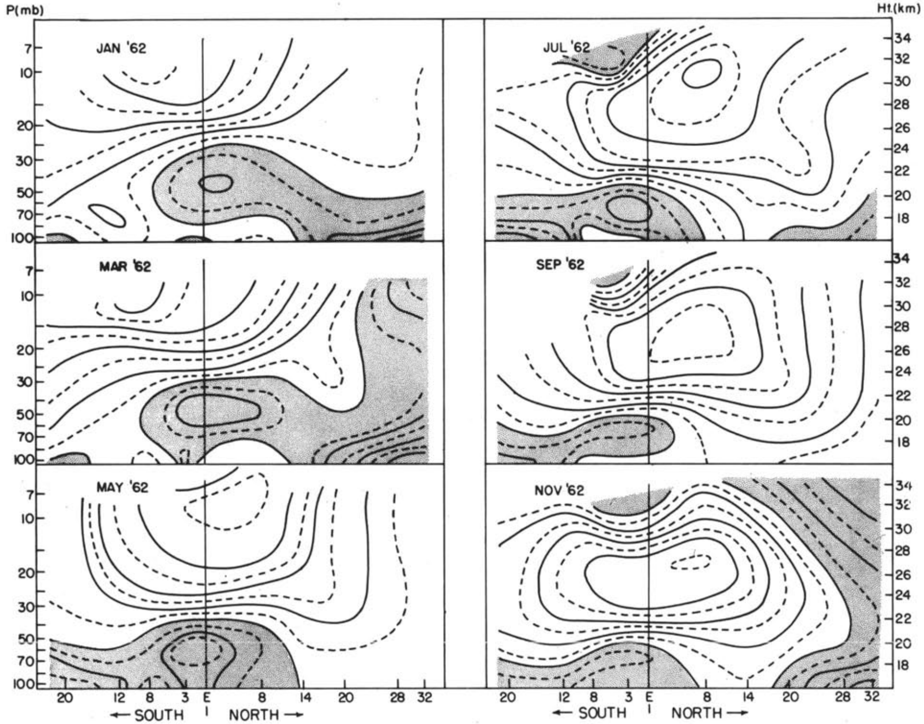


Figure 3.4(f) Meridional cross sections of zonal wind averaged around latitude circles for months indicated. Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies.

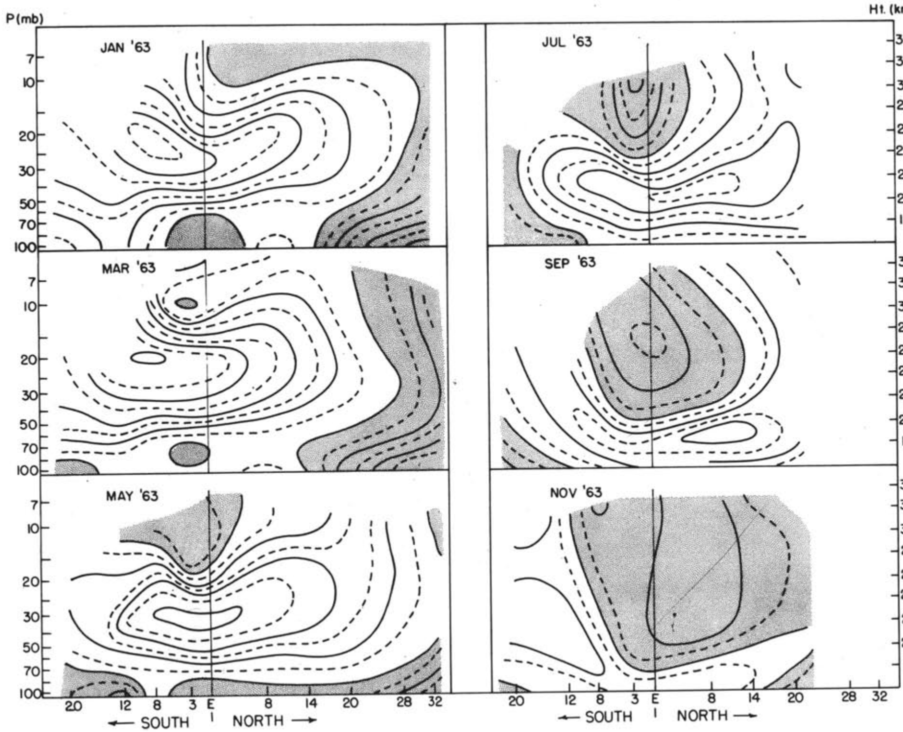


Figure 3.4(g) Meridional cross sections of zonal wind averaged around latitude circles for months indicated. Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies.

cross sections is suggestive of the motions of an extensive, axially symmetric vortex which is constantly adjusting itself in order to remain in equilibrium with its surroundings. This is a region where the non-symmetric disturbances represent truly small perturbations in a mean flow field which is undergoing change on a time scale much longer than the lifetime of the individual disturbances.

The remainder of this chapter will be devoted to examining the requirements which the global momentum and heat balances place upon this vortex, and the field of mean meridional motions which they induce.

Table 3.1. List of Symbols

λ = longitude in degrees

φ = latitude in degrees

r = the radial coordinate with origin at the center of the earth

z = the vertical coordinate

t = time

u = horizontal velocity component directed along latitude circles

v = horizontal velocity component directed along meridians

List of Symbols (cont)

w = vertical velocity component

Ω = the earth's angular velocity

a = the earth's radius

R = the gas constant

c_p = the specific heat of air at constant pressure

f = the Coriolis parameter ($2\Omega \sin\varphi$)

g = the acceleration of gravity

ρ = density

p = pressure

s = stability parameter = rate of increase of potential temperature with height

F = the frictional force per unit volume

H = time rate of temperature change due to radiative effects

$\overline{(\ )} = \frac{1}{t_2 - t_1}\int_{t_1}^{t_2} (\ ) dt$ = time average over one month

$(\ )' = (\ ) - \overline{(\ )}$ = deviation from time average

$[(\ )] = \frac{1}{2\pi}\int_0^{2\pi} (\ ) d\lambda$ = zonal average

$(\ )^* = (\ ) - [(\ )]$ = deviation from a zonal average

ω = time rate of change of pressure following a parcel

3.4 The Momentum Budget

(a) Derivation of the momentum balance equation

The equation governing the balance of westerly momentum may be derived from the zonal equation of motion and the equation of continuity in the following manner:

the zonal equation of motion,

$$\frac{du}{dt} = -\frac{1}{\rho r\cos\varphi}\frac{\partial p}{\partial\lambda} + fv + F_\lambda - 2\Omega\cos\varphi\, w + \frac{uv}{r}\tan\varphi - \frac{uw}{r} \qquad (1$$

may be expanded and multiplied through by density to give

$$\rho\frac{\partial u}{\partial t} = -\frac{1}{r\cos\varphi}\frac{\partial p}{\partial\lambda} + \rho f v + \rho F_\lambda - 2\Omega\cos\varphi\,\rho w + \frac{\rho u v}{r}\tan\varphi$$
$$-\frac{\rho u w}{r} - \frac{\rho v}{r}\frac{\partial u}{\partial\varphi} - \frac{\rho u}{r\cos\varphi}\frac{\partial u}{\partial\lambda} - \rho w\frac{\partial u}{\partial r} \qquad (2$$

The last five terms on the right may be further expanded in the forms

$$-\frac{\rho u w}{r} - \rho w\frac{\partial u}{\partial r} = -\frac{\rho w}{r}\frac{\partial}{\partial r} r u = -\frac{1}{r^3}\frac{\partial}{\partial r} r^3\rho u w + \frac{u}{r^2}\frac{\partial}{\partial r} r^2\rho w \qquad (3$$

$$\frac{\rho u v}{r}\tan\varphi - \frac{\rho v}{r}\frac{\partial u}{\partial\varphi} = -\frac{\rho v}{r\cos\varphi}\frac{\partial}{\partial\varphi} u\cos\varphi = -\frac{1}{r\cos^2\varphi}\frac{\partial}{\partial\varphi}\rho u v\cos^2\varphi + \frac{u}{r\cos\varphi}\frac{\partial\rho v\cos\varphi}{\partial\varphi} \qquad (3$$

$$-\frac{\rho u}{r\cos\varphi}\frac{\partial u}{\partial\lambda} = -\frac{1}{r\cos\varphi}\frac{\partial}{\partial\lambda}\left(\frac{1}{2}\rho u^2\right) + \frac{u}{r\cos\varphi}\frac{\partial\rho u}{\partial\lambda} \qquad (3$$

The last terms on the right in these three expressions, taken together, represent the divergence of the density weighted velocity vector in spherical coordinates, times the zonal wind. Making use of (3) together with the continuity equation, it is possible to write (2) in the form

$$\frac{\partial}{\partial t}\rho u = -\frac{1}{r\cos\varphi}\frac{\partial p}{\partial \lambda} + \rho f v + \rho F_\lambda - 2\Omega\cos\varphi\, w$$
$$-\frac{1}{r^3}\frac{\partial}{\partial r}r^3\rho u w - \frac{1}{r\cos^2\varphi}\frac{\partial}{\partial\varphi}\rho u v\cos^2\varphi - \frac{1}{r\cos\varphi}\frac{\partial}{\partial\lambda}\left(\frac{1}{2}\rho u^2\right) \tag{4}$$

At this point several useful simplifications may be introduced.

(1) The radial coordinate r may be regarded as a constant a, equal to the earth's radius, wherever it is used as a multiplicative factor. Where used in a differential operator it may be replaced by z, the height above the earth's surface.

(2) The variations of density in space and time at a given level are much smaller than variations in the wind components. Therefore, ρ may be treated as a function of z only. With these assumptions, (4) becomes

$$\rho\frac{\partial u}{\partial t} = -\frac{1}{a\cos\varphi}\frac{\partial p}{\partial\lambda} + \rho f v + \rho F_\lambda - 2\Omega\cos\varphi\, w$$
$$-\frac{\partial}{\partial z}\rho u w - \frac{\rho}{a\cos^2\varphi}\frac{\partial}{\partial\varphi}uv\cos^2\varphi - \frac{\rho}{a\cos\varphi}\frac{\partial}{\partial\lambda}\frac{u^2}{2} \tag{5}$$

Next, the averaging processes, first over a month and then around the latitude circle, may be applied to (5) which then becomes, after dividing through by the density,

$$\frac{\partial[\bar{u}]}{\partial t} = f[\bar{v}] + [\bar{F}_\lambda] - 2\Omega\cos\varphi[\bar{w}] - \frac{1}{\rho}\frac{\partial}{\partial z}\rho\left\{[\bar{u}][\bar{w}] + [\bar{u}^*\bar{w}^*] + [\overline{u'w'}]\right\}$$
$$-\frac{1}{a\cos^2\varphi}\frac{\partial}{\partial\varphi}\cos^2\varphi\left\{[\bar{u}][\bar{v}] + [\bar{u}^*\bar{v}^*] + [\overline{u'v'}]\right\} \tag{6}$$

the terms with derivatives with respect to longitude having vanished in the zonal average.

The first term in each of the braces represents the effects of mean meridional motions. These may be expanded and combined in the form

$$-[\bar{w}]\frac{\partial[\bar{u}]}{\partial z}-\frac{[\bar{v}]}{a\cos\varphi}\frac{\partial[\bar{u}]}{\partial\varphi}-\frac{[\bar{u}]}{\rho}\left\{\frac{\partial}{\partial z}\rho[\bar{w}]-\frac{1}{a\cos\varphi}\frac{\partial}{\partial\varphi}\rho[\bar{v}]\cos\varphi\right\}$$

where the expression in braces vanishes from continuity considerations.

The frictional term represents the effects of all eddies of too small a time scale to be measurable with once daily observations. Since the momentum balance will be applied to latitudinal strips whose meridional dimension is two orders of magnitude larger than their vertical dimension and since the disturbances which are responsible for eddy viscosity are not likely to be as quasi-horizontal as this, it will be assumed that the main effect of friction is to exchange zonal momentum in the vertical by means of a correlation between zonal and vertical motions. This is basically the same mechanism by which momentum is exchanged in the vertical by synoptic scale eddies. Since neither term can be measured directly it is convenient to combine the two effects in the vertical flux term which will henceforth be understood to represent the effects of both synoptic and subsynoptic scale eddies.

The zonally averaged equation governing the balance of westerly momentum may then be written as

$$\frac{\partial[\bar{u}]}{\partial t} = -\frac{1}{a\cos^2\varphi}\frac{\partial}{\partial\varphi}\cos^2\varphi\left\{[\overline{u'v'}] + [\bar{u}^*\bar{v}^*]\right\}^{(1)}$$

$$+\left(f - \frac{1}{a\cos\varphi}\frac{\partial}{\partial\varphi}[\bar{u}]\cos\varphi\right)[\bar{v}]^{(2)} - \frac{\partial[\bar{u}]}{\partial z}[\bar{w}]^{(3)} \qquad (7)$$

$$-2\Omega\cos\varphi[\bar{w}]^{(4)} - \frac{1}{\rho}\frac{\partial}{\partial z}\rho\left\{[\overline{u'w'}] + [\bar{u}^*\bar{w}^*]\right\}^{(5)}$$

where the first term on the right represents the convergence of the horizontal transport of westerly momentum by transient and standing eddies, respectively, the second, third and fourth terms give the effects of the mean motions, and the final term gives the convergence in the vertical transport of westerly momentum by all scales of eddies.

(b) Simplifications resulting from scaling considerations

Before discussing the momentum budget in general it is convenient to assess the importance of the last two terms in the above equation. The fourth term on the right hand side represents the advection of the earth's angular momentum by the mean vertical motions. It can be seen by comparing typical magnitudes of the coefficients of $[\bar{w}]$ in terms (3) and (4) that this effect is almost two orders smaller than that of the previous term and thus it can be neglected.

Term (5) represents the divergence of the vertical flux of momentum by all scales of eddies and includes frictional effects. Dickinson

(1962) has computed the contribution of synoptic scale eddies to this term, using adiabatically computed vertical velocities at middle latitudes. It was found that the vertical flux divergences associated with this scale of motion are about half an order of magnitude smaller than the horizontal divergences of term (1), and that the momentum transports are, in general up gradient. It is possible that less organized, smaller scale motions might counteract this effect by transporting momentum down the gradient.

At very low latitudes the sections show significant vertical transports of momentum associated with the downward propagation of successive easterly and westerly wind regimes. However there are several difficulties involved in attempting to attribute this exchange to a vertical diffusion process.

(1) The very nature of the diffusion process implies a change in strength of the gradients down or up which the mixing takes place. It is evident from the time-height section for 3^{o}S in Fig. 3.1(b) that regions of strong vertical zonal wind gradient propagate downwards from 20 mb to 50 mb without any substantial diminution of strength. Below 50 mb, gradients tend to diminish with further downward propagation, which suggests that down gradient diffusion may be of importance there.

(2) Measurements of the vertical movements of trace substances in the tropical stratosphere by Friend et al (1961), suggest an eddy diffusion coefficient in the order of 10^{3} $cm^{2}sec^{-1}$. If this value is applied to the

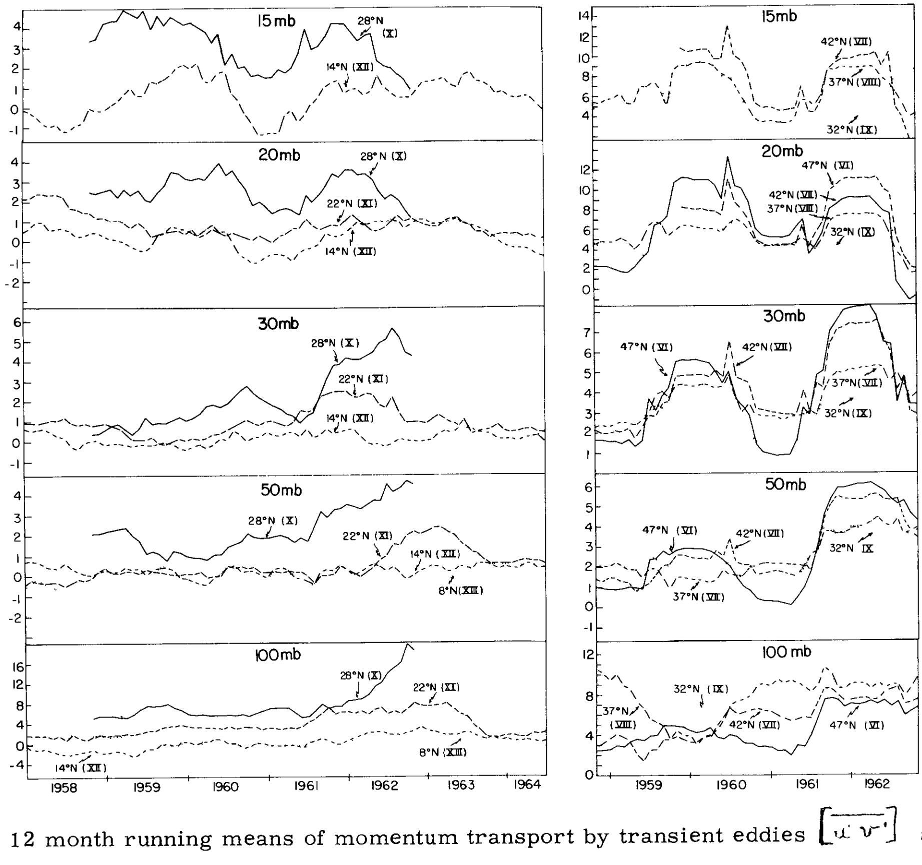
diffusion of momentum it is found to be almost an order of magnitude too small to account for the observed zonal accelerations at 50 mb and above.

Thus, with the possible exception of the region below 50 mb it appears reasonable to neglect the role of eddy processes in the vertical exchange of momentum associated with the long period wind fluctuations the tropics. With this assumption the right hand side of the momentum equation reduces to the first three terms. Term (1) represents the horizontal divergence of eddy momentum transports. Reed et al. (1961) has shown that this is the only mechanism capable of introducing westerly momentum into the equatorial region. Similarly it is the only mechanism capable of removing westerly momentum from the tropical region as a whole. It must therefore be responsible for all changes in vertically integrated westerly momentum which are symmetric about the equator. This last condition excludes the annual cycle.

(c) Transient eddies

Tucker (1964, 1965) has already presented evidence of variations in the transport of westerly momentum by transient eddies. The large data sample available in the present study provided a means of describing this phenomenon in more detail. Figure 3.5 shows 12 month running means of $[\overline{u'v'}]$ for various latitude bands and levels. The following features are to be noted:

1. There do indeed appear to be long period variations in the eddy



e 3.5 12 month running means of momentum transport by transient eddies $[\overline{u' v'}]$ avera around latitude circles as indicated for five levels between 100 and 15 mb in the norther hemisphere. Units are in $m^2\ sec^{-2}$. Means are centered on the dates indicated. Dec momentum transports and their latitudinal divergence above 30 mb in the tropics and above 50 mb at higher latitudes. These are real features which show up in many basically independent sets of data at many different latitudes and levels.

2. Below 30 mb the data give little or no evidence for any definite year to year variation in the momentum transports at low latitudes.

3. The oscillation in momentum transports is practically simultaneous at all levels and latitudes where it appears.

These results will be discussed in more detail in section 4. 2a.

Because of its relatively large inertia the tropical stratosphere responds to the momentum transports accumulated over a period of months rather than to the individual disturbances which produce the transports. This unique property of this region of the atmosphere is responsible for the smoothness of the time-height sections. At 20 mb and above, the variations in the convergence of the transport of westerly momentum are almost of sufficient size* and occur at the proper times to account for the observed long period zonal accelerations. However, this mechanism

*From Fig. 3. 4 it appears that the momentum transports increase from small values at the equator to alternating positive and negative values with an amplitude in the order of 2 $m^2 \sec^{-2}$ at 20°N. This produces convergences and divergences in the order of 10^{-6} m $\sec^{-2}$. Typical zonal accelerations as deduced from Fig. 3. 1b are in the order of 5 m $\sec^{-1}$ mo^{-1} and extreme values reach 10 m sec mo^{-1} during late 1963. In units comparable with the flux divergences, these correspond to 2 and 4 x 10^{-6} m $\sec^{-2}$, respectively.

cannot explain the propagation of the same regimes into the lower stratosphere, since the observed transports fail to exhibit any year to year differences in the tropics below 30 mb and there is no evidence of the phase shift in the vertical that is observed in the zonal winds. Thus it appears that some type of mean motion is also necessary to explain the observed wind field.

(d) Standing eddies

It is not possible to make a direct estimate of the contribution of standing eddies to the transport of momentum at low latitudes because the station distribution is far too sparse to permit any representation of the monthly mean winds around an entire latitude circle. However there is some indirect evidence regarding their importance as compared with transient eddies.

As seen in Table 3. 2, typical station values of $\bar{v}$ for a given month rarely deviate from zero by more than 2 m sec^{-1} at these latitudes and more typical values are less than a meter per second. This suggests that the standard deviation of the meridional component in the standing eddies is smaller, by at least a factor of two, than it is in the transient eddies. The time-height sections provide some idea of the variability of the monthly mean zonal component along a latitude circle. The ease with which data from different longitudes could be pieced together, and the apparent reliability of the analyses suggests a standard deviation of

Table 3.2 Mean monthly values of the meridional wind component during 1964 for selected tropical stations. Units in m sec^{-1}.

	Jan	Apr	July	Oct
Balboa				
70 mb	0.1	0.3	0.3	0.0
50	0.8	0.2	0.2	-0.3
30	-1.2	-0.3	0.6	-0.4
20	0.2	0.4	-0.2	
10	-0.6	0.8	-0.4	
Canton Is.				
70 mb	0.2	-1.1	-0.5	0.3
50	-0.1	0.0	-0.4	-0.4
30	-0.2	0.4	-0.2	-0.6
20	-0.9	-0.3	1.3	0.2
10	-0.3	-0.7	-0.2	-0.9
Clark AFB, Philippines				
70 mb	0.4	-1.7	-2.2	-1.1
50	0.1	-0.7	-1.6	-0.8
30	0.4	-0.4	-1.0	-1.1
20	-0.4	-0.9	-2.1	-1.1
10	-0.5	-1.8	-2.4	-1.8

considerably less than the value of 5 m sec^{-1} which is typical of transient eddies. If these estimates are reasonable, it would seem that standing eddies are of secondary importance in the transport of momentum unless, for some reason, there is a very high correlation between $\bar{u}$ and $\bar{v}$.

At higher latitudes in the winter hemisphere the situation is quite the reverse; it will be necessary to consider standing eddies again in the discussion of the hemispheric circulations in the next chapter. (Sec. 4. 2b).

(e) Mean motions

Before considering the terms in the momentum balance which involve mean motions, it would be well to consider whether there is any limit on the size of mean motions which this region of the atmosphere can tolerate. It is apparent from Fig. 3. 1 that typical zonal accelerations in this region range from 5 m sec^{-1} mo^{-1} at the equator to perhaps three times that value at 20^{o} latitude; in more standard units, a range of 2-6 x 10 m sec^{-2}. The momentum flux data in Fig. 3. 5 suggest divergences ranging from 1 m^{2} sec^{-2} per degree of latitude near the equator to perhaps three times that value at 20^{o}N. In standard units, these values correspond to 1-3 x 10^{-6} m sec^{-2}, which is slightly smaller, but roughly the same order of magnitude as the zonal accelerations. If equation (7) is to be satisfied, the mean motion terms cannot be much larger than this unless they tend to compensate each other. Since continuity must be satisfied by the horizontal and vertical components of the mean motions, it is unlikely that such compensation will be a general rule. The Coriolis parameter is about 2.5×10^{-5} sec^{-1} at 10^{o} and 5×10^{-5} sec^{-1} at 20^{o} of latitude, which limits the mean meridional motions to values in the order of a tenth of a meter per second. Because of continuity, this also implies certain restrictions on vertical motions, which will be discussed below.

The second term on the right hand side of equation (7) represents the effects of mean meridional motions. Dickinson (1962) has pointed out that this is the only term which could be large enough to account for the seasonal wind reversals evident at subtropical latitudes in Figs. 3.1 and 3.4. (He estimated that a mean drift of about 5 cm sec^{-1} from the spring hemisphere into the autumn hemisphere could bring about the observed accelerations. This would amount to a displacement amplitude of the order of a few degrees of latitude over the course of a year). However, it can be shown that within a degree of the equator, where the Coriolis parameter and the latitudinal gradients of the zonal wind are both very small, the magnitude of this term is small compared with the observed zonal wind accelerations. At 1^{o} of latitude $f \sim 2.5 \times 10^{-6}$ sec^{-1} and $[\bar{v}] \sim 0.1$ m sec^{-1}. Thus the Coriolis accelerations are in the order of $.25 \times 10^{-6}$ m sec^{-1}, which is an order of magnitude smaller than the observed accelerations. Fig. 3.4 suggests that the meridional gradient of zonal wind is in the order of 1 m sec^{-1} per degree of latitude near the equator or 10^{-6} sec^{-1} which is even smaller than the Coriolis parameter at 1^{o}. Thus by the process of elimination, it appears that the third term, representing the advection of momentum by mean vertical motions must be responsible for the downward propagation of the wind regimes at the equator. This is essentially the same result as Tucker (1964) arrived at in his more theoretical formulation. This means that the slopes of the isotachs in the time-height section for 3^{o}S (Fig. 3.1(b)) should give a good representation of the vertical motion field over the equator between about 20 and 50 mb. An inspection of Fig. 3.1(b) suggests a mean downward drift of a little less than 1 km/mo or about .03 cm sec^{-1}. This method of inferring vertical motions is not applicable more than a degree or two away from the equator where mean meridional motions are also capable of effecting an exchange of momentum in the vertical.

The steeper descent rate of westerlies than easterlies at the equator, as mentioned in the previous section, implies that the mean downward motion is enhanced where the vertical wind shear is positive, and diminished where it is negative. It follows from continuity that the associated meridional motions would be adjusted by this differential descent rate in such a way as to effect a vertical exchange of momentum in the same sense as the vertical motions. For example, if at the equator westerlies are replacing easterlies at some level, and downward motion is stronger than usual there, then at a slightly higher latitude, the flow is more equatorward than usual above this level, and more poleward than usual below. Coriolis torques are generating easterlies above and westerlies below, which has the same effect as a downward transport of westerly momentum at the higher latitude. Such a mechanism seems more likely to be responsible for the downward propagation of the the oscillation at subtropical latitudes than the advection of momentum by a mean downward drift, since as Tucker (1964) has shown, the mean meridional motions associated with a uniform mean downward drift within 20° of the equator of the size thus required would be in the order of 50 cm sec^{-1} at 20°. This would be large enough to wreak havoc with the momentum budget at these latitudes.

3.5 The Temperature Field and its Relation to the Wind Field

(a) Geostrophic balance at low latitudes

It has long been a recognized fact that the geostrophic equation is not valid for synoptic scale disturbances at low latitudes. However, the motions under investigation may be geostrophic to within a very short distance of the equator because of their very long time scale. To investigate this possibility it is necessary to use the equation of motion for the meridional component.

$$\frac{dv}{dt} = -\frac{1}{\rho a}\frac{\partial p}{\partial \varphi} - u \sin\varphi \left(2\Omega + \frac{u}{a\cos\varphi}\right) - \frac{wv}{a} + F_\lambda \tag{8}$$

where a $\sim$ 6.4 x 10^6m

$2\Omega \sim$ 1.5 x 10^{-4} sec^{-1}

u $\sim$ 10 m sec^{-1}

For these low latitudes, $\cos\varphi \sim 1$ and $\sin\varphi \sim y/a$ where y is the distance from the equator.

It is readily seen that $2\Omega \gg u/a\cos\varphi$, and therefore the second term in parentheses may be dropped. The equation is then averaged, first over a month and then around a latitude circle. With the result

$$\frac{\partial[\bar{v}]}{\partial t} = -\frac{1}{a}\left[\overline{\frac{1}{\rho}\frac{\partial p}{\partial \varphi}}\right] - f[\bar{u}] - \frac{1}{a}\left\{[\bar{w}][\bar{v}] + [\bar{w}^*\bar{v}^*] + [\overline{w'v'}]\right\} \tag{9}$$

Now an order of magnitude estimate can be placed on each of the terms except the pressure gradient force:

$d[\bar{v}]/dt$ may be approximated by $\partial[\bar{v}]/\partial t$ for these very slow motions. In Sec. 3.4e, it was shown that typical values of $[\bar{v}]$ are in the order of 10^{-1} m sec^{-1} or less and these variations occur on a time scale of several months, or 10^7 seconds. Accordingly, the accelerations should be in the order of 10^{-8} m sec^{-2}.

The Coriolis term evaluated at a distance of 1 km from the equator amounts to about 20 x 10^{-8} m sec^{-2}.

The wv correlation terms include the major effects of friction if all scales of eddies are taken into account. Assuming values of 0.3×10^{-2} m sec^{-1} for $[\bar{w}]$, and 10^{-1} m sec^{-1} for $[\bar{v}]$, as deduced in Sec. 3.4e, the mean term is smaller than 10^{-11} m sec^{-2}. Previous estimates in Sec. 3.4d indicate that the transient eddy term should be the larger of the two remaining terms. Table 2.1 suggests that 3 m sec^{-1} is an appropriate value for the standard deviation of v, and 10^{-2} m sec^{-1} should be a liberal estimate for the standard deviation of w, according to Newell (1963). A correlation coefficient of 0.2 would lead to a value in the order of 10^{-9} m sec^{-2} for this term.

It is evident that the pressure gradient force is the only term large enough to balance the Coriolis force and hence the geostrophic equation is valid to well within 1 km of the equator for these long period, zonally symmetric motions. This is indicative of a strong relation between the zonally averaged wind and temperature fields in the tropical stratosphere. Reed (1962, 1964b) noted long period variations in the temperature field and succeeded in relating them geostrophically to the wind field. He has noted that because wind observations provide a relatively greater resolution of the long period variations, the best available estimate of the temperature variations is that deduced from the wind field using the thermal wind equation.

An important question arises as to the relation between the wind and temperature fields: namely, are the variations in one field indirectly responsible for the variations in the other, through the action of mean meridional motions? An answer to this question is implicit in any comprehensive model of these phenomena. For instance, the models of Staley (1963), Probert Jones (1964) and Lindzen (1966) tacitly assume that the temperature field plays the active role, as does any model which calls upon radiative effects to produce thermal variations in situ. The absence of theories which ascribe to the wind field the active role is probably due more to the difficulty in treating the eddy fluxes in the momentum equation than to any well founded belief that the wind field is of secondary importance. Thus far there is no evidence of any variability in solar output strong enough, and of the proper period to account for the observed temperature variations. On the other hand, the presence of long period variations in eddy fluxes of momentum is an observed fact. It will presently be shown how changes in the wind field resulting from these fluxes give rise to mean meridional motions* which maintain the temperature field in geostrophic equilibrium with the wind field, thus explaining the observed temperature variations.

*Mean meridional motions, by continuity considerations imply corresponding vertical motions.

(b) The heat balance

It is most convenient to derive the heat balance equation in a system with pressure as the vertical coordinate. In this system, the derivation is similar to that of the momentum balance equation, and therefore it will be omitted. The derived equation as adapted from Saltzmann (1961) is of the form

$$\frac{\partial [\bar{T}]}{\partial t} = -\frac{[\bar{v}]}{a}\frac{\partial [\bar{T}]}{\partial \varphi} - [\bar{\omega}]\left(\frac{\partial [\bar{T}]}{\partial p} - \frac{R[\bar{T}]}{c_p p}\right) - \frac{1}{a\cos\varphi}\frac{\partial}{\partial \varphi}\left\{[\overline{v'T'}] + [\bar{v}^* \bar{T}^*]\right\} - \frac{\partial}{\partial p}\left\{[\bar{\omega}^* \bar{T}^*] + [\overline{\omega' T'}]\right\} + \frac{R}{c_p p}\left\{[\overline{\omega' T'}] + [\bar{\omega}^* \bar{T}^*]\right\} + H \qquad (10)$$

where the symbols are as listed in Table 3.1.

Since it is to be used only for a rough order of magnitude comparison of the various terms, equation (10) can be expressed with height as a vertical coordinate simply by replacing ω by $-\rho g w$ wherever it occurs and $\partial/\partial p$ by $-\frac{1}{\rho g}\,\partial/\partial z$;

$$\frac{\partial [\bar{T}]}{\partial t} = -\frac{[\bar{v}]}{a}\frac{\partial [\bar{T}]}{\partial \varphi} - \lambda[\bar{w}] - \frac{1}{a\cos\varphi}\frac{\partial}{\partial \varphi}\left\{[\overline{v'T'}] + [\bar{v}^* \bar{T}^*]\right\} - \frac{1}{\rho}\frac{\partial}{\partial z}\rho\left\{[\overline{w'T'}] + [\bar{w}^* \bar{T}^*]\right\} - \frac{g}{c_p[\bar{T}]}\left\{[\overline{w'T'}] + [\bar{w}^* \bar{T}^*]\right\} \qquad (11)$$

where λ expresses the rate of increase of potential temperature with height.

Of the first two terms, which represent the advective effects of the mean motions, the second is several orders of magnitude larger if the following are assumed as typical values in the region:

$$1/a \; \partial T/\partial \varphi \sim 2^{\circ}\mathrm{C}/1000 \text{ km}$$

$$s \sim 13^{\circ}\mathrm{C}/\mathrm{km}$$

$$[\bar{v}] \sim 10^{-1} \text{ m sec}^{-1}$$

$$[\bar{w}] \sim 10^{-4} \text{ m sec}^{-1}$$

These estimates suggest heating rates in the order of tenths of a degree per day resulting from mean vertical motions.

Data on the magnitude of the eddy heat transports will be presented in the next chapter. For an order of magnitude estimate it will be noted from Fig. 4. 2 that the four year mean of the heat transport increases from near zero at the equator to about $0.5^{\circ}\mathrm{C}$ m sec^{-1} at $20^{\circ}\mathrm{N}$. This eddy flux divergence corresponds to a cooling rate in the order of a few hundredths of a degree Centigrade per day.

To determine the importance of the terms involving vertical eddy fluxes it is necessary to make some estimate of the variability of temperature and vertical motion. The results of Sec. 3. 4d suggest that the temporal variability is probably a good estimate of the total. Using Newell's (1963) value of 10^{-2} m sec^{-1} as an upper limit of the standard deviation of the vertical wind component, a value of $2^{\circ}\mathrm{C}$ for the standard deviation of temperature, as suggested by Table 2. 1, and a correlation coefficient of 0.2, we arrive at an estimate of 0.4×10^{-2} °C m sec^{-1} for $[\omega' T']$. Assuming a vertical scale of 10 km, this leads to a heating rate in the order of hundredths of a degree per day for the divergence of heat flux. The coefficient of the vertical flux term, $\frac{g}{c_p[T]}$, is roughly 0.5×10^{-4} m^{-1}, which suggests that this term is somewhat smaller than the divergence term.

The most recent estimate of the diabatic heating rate, H, for this region was made by Kennedy (1964). The heating rates which he computed vary considerably with altitude, season and cloudiness, but are generally in the order of tenths of a degree per day.

The only remaining term, the local time derivative of the temperature, is very small, since the temperature range is only a few degrees and the time scale is in the order of months.

(c) The maintenance of geostrophic equilibrium

Thus thermal equilibrium is to a large extent accomplished by a balance between radiation and vertical motions, with the other terms exerting only a small influence which can be neglected in this qualitative discussion.

To show how geostrophic equilibrium is maintained between the wind and temperature fields let us consider a temperature field in radiative equilibrium in an atmosphere with only zonal, geostrophic motions.

Let it be supposed that eddy fluxes produce a disturbance in the zonal wind field. The temperature field will immediately begin to adjust to a new equilibrium state, through the action of forced mean meridional motions. For simplicity, we will neglect any effects which the vertical motions might have in altering the distribution of the important radiative constituents of the atmosphere.* In the absence of such effects, the radiativè heating rate should increase in the adiabatically cooled regions and decrease in the adiabatically warmed regions so as to resist deviations from radiative equilibrium. If it were not for continued vertical motions, these changes in heating rates would eventually bring the temperature field back into radiative equilibrium. The speed with which this readjustment would occur could be described by the radiative relaxation time, i.e., the time necessary for a disturbance to be attenuated to 1/e of its original value. (Manabe and Strickler's (1964) numerical experiments with the radiative transfer equation suggest a value in the neighborhood of 20 days for this parameter). In reality, of course, mean vertical motions will maintain the temperature disturbance so long as the disturbance in the wind field exists. The effect of radiative relaxation will be to require vertical motions larger than would be needed to establish the geostrophic temperature gradient in an adiabatic atmosphere. Moreover, the need for these motions will not cease once

*In the real atmosphere, the effect of the motions upon the ozone and water vapor distributions and their radiative properties may not be negligible.

the gradient has become geostrophic; it exists so long as there is a temperature gradient to maintain against radiative relaxation. In fact, the order of magnitude estimate of the terms in the heat balance equation (11) suggests that the motions required to establish the gradient are negligible compared those necessary to counteract the dissipative effect of radiation, i.e., $H \gg \partial[\bar{T}]/\partial t$

(d) The energy cycle

From the standpoint of energetics it is apparent that the zonal available potential energy of the disturbance is continually being destroyed by radiation, on the time scale of the radiative relaxation time. The actual amount of zonal available potential energy in the disturbance is kept from decreasing by mean meridional motions which convert enough energy from the zonal wind field to supply what is lost by radiation. This results in a net drain on the zonal kinetic energy of the disturbance. The effect on the wind field is very slow to be felt because in this region of the atmosphere the kinetic energy is larger than the available potential energy by about a factor of 40*, and hence the dissipative time scale is

*Lorenz (1955) showed that for the atmosphere as a whole the ratio of kinetic energy to available potential energy, K/A, is about 1/10. When the tropical stratosphere alone is considered, the situation is quite different, for two reasons:

(a) The static stability, λ, is about four times as large as that which Lorenz used for the atmosphere as a whole.

(b) The meridional gradients of the zonally averaged temperature are only about a tenth the size which Lorenz estimated for the entire atmosphere.

Hence the available potential energy, which is proportional to the variance of temperature divided by static stability is only 1/400 the average value for the whole atmosphere. The kinetic energy of the region, on the other hand, should be typical of the global value. Hence the ratio 40:1 for K/A.

in the order of 3 years. (40 relaxation times).

In retrospect, the difficulties in trying to explain the observed wind variations as resulting from variations in solar heating of the region are obvious. In the first place it is difficult to see how a small change in heating could create any substantial amount of available potential energy, which requires not heating, but a gradient of heating. Even if this were possible, the zonal available potential energy of the temperature field is but a small fraction of the kinetic energy of the wind field and thus differential heating would have to generate many times the amount of energy required to account for the temperature variations. The energy changes resulting from variations in the momentum transport are many times larger than those due to any known fluctuations in radiative heating.

3.6 The Field of Vertical Motions

The assumption of a mean downward drift at the equator leads to the notion of a radiative heat sink in that region. Otherwise it would be impossible to account for the fact that the equator is colder than the subtropics at these levels. Because of uncertainties in dealing with the radiative transfer equations which are only now being dealt with (see Rodgers and Walshaw (1966)) and the lack of data available on the distribution of ozone and water vapor at low latitudes it is difficult to either support or refute this hypothesis from radiative considerations at present. (Only at the tropopause itself can a mean downward motion be ruled out from radiative considerations). Considering that our present understanding of the momentum budget of this region is at least as good as that of the heat budget, it would seem no less direct to infer the vertical motion field from the momentum budget than to infer the meridional motion field from the heat budget as was done by Murgatroyd and Singleton (1961). A mean downward drift in the order of .03 cm sec^{-1}, based on the descent rate of the wind regimes, would lead to a required radiative cooling rate in the order of 0.3°C per day, which is of roughly the same magnitude as that computed by Kennedy (1964), but of opposite sign.

From geostrophic considerations, warm temperatures are required in the tropics relative to the subtropics in the region where the wind shear is positive, i.e., where westerlies are propagating downward into easterlies. Since the temperature is determined by a balance between radiation and adiabatic heating or cooling due to vertical motion, an increase in downward vertical motion is required in order to maintain a warm temperature in this region against radiative relaxation. The model proposed by Reed (1964b), which invoked a system of downward propagating mean meridional cells is based on essentially the same physical mechanism.

It is difficult to assess the magnitude of the vertical motions thus caused because of the lack of more reliable information on radiative relaxation times. For an order of magnitude estimate, let it be assumed that in the long term mean there is a net downward motion of .03 cm sec^{-1} which is balancing a cooling rate of 0.3°C per day. To maintain a temperature perturbation of 2°C from this mean state against a radiative relaxation time of 20 days this would require a corresponding change in vertical motion of about .01 cm sec^{-1}, which is consistent with Reed's (1964b) results. Under these conditions, westerlies would propagate downward at approximately twice the speed of easterlies, and this is approximately what is observed. Thus it seems likely that the peculiar shape of the wind regimes in the time-height plane arises from the requirements of the heat balance. Furthermore, as stated in Sec. 3.4e, the meridional motions arising from the same vertical motion field may be responsible for the broad latitudinal extent of the wind variations.

The vertical motion field suggested by this analysis is similar to the models of both Tucker (1964) and Reed (1964b); in fact, it is a superposition of the two - a steady, descending current over the equator modulated by a weaker secondary cell which propagates downward with the zonal wind regimes. The fact that westerlies are observed to propagate downward more rapidly than easterlies lends support to this hypothesis.

3.7 Concluding Remarks

The combination of low latitude with high static stability is a unique feature of the tropical stratosphere. The former condition severely restricts the amplitude of geostrophic temperature fluctuations, while the latter still further reduces the amount of available potential energy present in the temperature field. This combination of circumstances results in a situation where the greater part of the available energy of the region resides in the mean zonal kinetic energy of the wind field. This is in marked contrast to the atmosphere as a whole, where the kinetic energy constitutes only a small fraction of the available energy.

It appears that the long period zonally symmetric disturbances derive their energy from the eddy kinetic energy of the circulations at higher latitudes. As evidence of this, it can be seen by comparing Figs. 3.4 and 3.5 that above 30 mb during the winter months (it will be shown in Sec. 4.2 that this is where and when the important eddy fluxes take place) the momentum transports are usually countergradient. Mean meridional motions convert some of this energy into zonal available potential energy in order to maintain the latter against radiative relaxation. This chain of cause and effect is summarized in Fig. 3.6.

Although many of the terms in the momentum and heat balance equations, particularly those pertaining to radiation are known only to

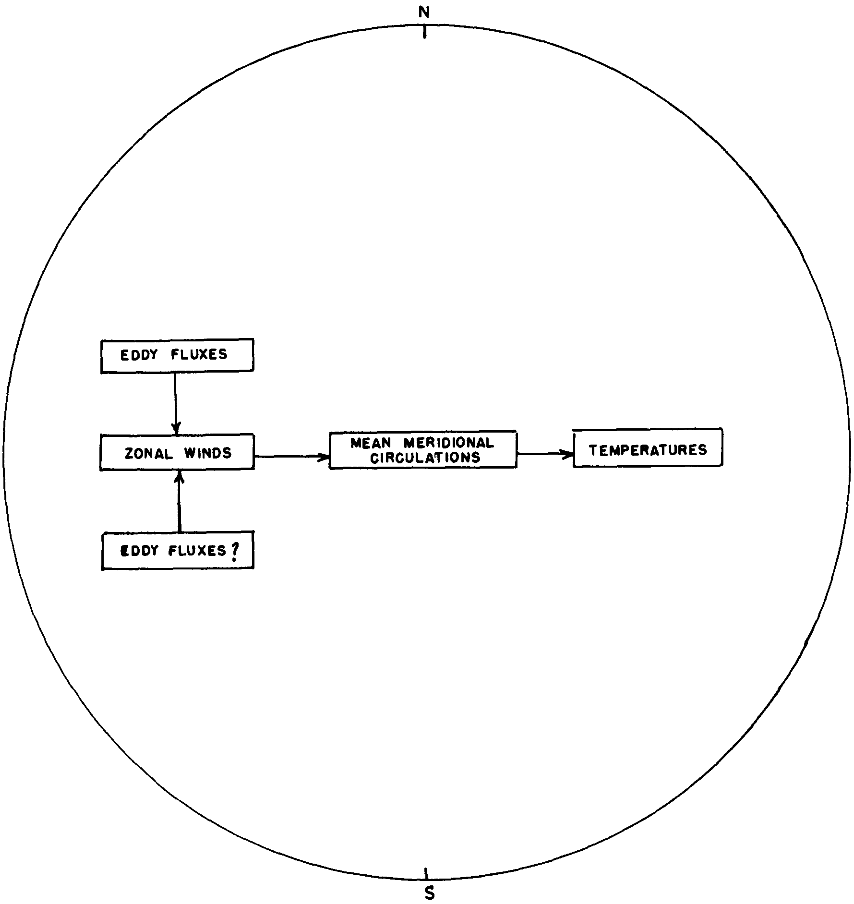

N
EDDY FLUXES
ZONAL WINDS
MEAN MERIDIONAL CIRCULATIONS
TEMPERATURES
EDDY FLUXES ?
S

Figure 3.6

within an order of magnitude, it would be difficult to conceive of an alternate description of the mechanics of the system that is internally consistent.

The treatment of the balance requirements in this chapter has purposely been kept as simple as possible in order to allow a clear qualitative understanding of the dynamical processes involved. It should not be difficult to formulate a more sophisticated model in which many of the smaller terms that have been neglected in the present treatment could be retained. The momentum and heat balance equations together with the thermal wind equation for the zonal component and the continuity equation comprise a system which can be solved for the variables $[\bar{u}]$, $[\bar{v}]$, $[\bar{w}]$ and $[\bar{T}]$, given appropriate boundary conditions and realistic specifications of the eddy fluxes, radiation and friction. Eliassen (1950) and Kuo (1956) have shown how this system of equations can be solved for the mean meridional motions and the time derivatives of the zonal wind and temperature fields. The simplified analysis in this chapter has suggested what might be appropriate values for radiation and friction. The former may be dealt within terms of a heat sink above 50 mb over the equator, and a relaxation time for disturbances from the mean state, while the latter may perhaps be neglected, at least in a first approximation. The only remaining quantity to be specified would then be the eddy fluxes. These could, of course, be modeled after actual data such as that presented in this chapter. Such a treatment would be of value in studying the effect of a given forcing function on the mean wind field, but it would fail to yield any insight into the ultimate cause of the wind and temperature variations in question. For that purpose it is necessary to know the cause of the long period variations in the momentum fluxes themselves. The next chapter is devoted to a study of the circulations at middle latitudes, which give rise to the momentum fluxes.

In a sense, the present chapter has been primarily concerned with the question of the effects produced by the eddy circulations on the mean wind and temperature fields. One of the purposes of the next chapter will be to determine what effects, if any, the mean fields have upon the eddy circulations.

CHAPTER IV

LONG PERIOD VARIATIONS IN THE HEMISPHERIC CIRCULATIONS OF THE STRATOSPHERE

Because of the marked year to year variations in the momentum transports at middle latitudes evident in Fig. 3.5, the hemispheric circulations of the stratosphere must be considered as an important link in the chain of cause and effect which gives rise to the so called biennial oscillation. Not only are these circulations involved in this phenomenon; they apparently are responsible for the effects observed in the tropics. The purpose of this chapter is to examine the nature of the long period variations in those aspects of the middle latitude circulations which have a direct effect on the energy budget of the tropics. Of special interest will be the question of cause and effect; i.e., "Are these long period variations in the momentum fluxes inherent characteristics of the stratospheric circulation, or are they reflections of variations in some still more basic quantity?" To begin with, a brief discussion of the seasonal behavior of the hemispheric circulations will be helpful in understanding the longer period variations.

4.1 The Seasonal Circulations

(a) Summer

Except at very low latitudes the summer and winter circulations in the stratosphere are dramatically different. The summer monsoon is marked by a gentle, undisturbed easterly flow above 50 mb at all latitudes. Warm temperatures over the pole give rise to an anticyclone which gradually grows in intensity with height above 50 mb. The air flows around this polar anticyclone in almost perfect circles so that the wind is very steady and uniform. Table 4.1 shows that above 100 mb, typical temporal standard deviations are 3 m sec^{-1} for the zonal component and 2 m sec^{-1} for the meridional component, which is even smaller than in the tropics. The four year mean values of the heat and momentum fluxes during the summer months are too small to even establish a prevailing direction of the transport at and above 50 mb. This complete lack of correlation suggests that the wind measurements (except for the mean zonal flow) during this season may be near the instrumental and mesoscale noise level. In any case, the eddy fluxes are negligible. Below 50 mb there are still some remnants of the westerly flow associated with the jet stream, which are mainly confined to high latitudes. At lower latitudes, at least in the northern hemisphere, the land-sea influence is felt, the most spectacular feature of which is the Tibetan anticyclone and the

Table 4.1. Transient eddy statistics at 42°N. Four year means (6/58 - 1/62).

	σu	σv	σT	$[\overline{u'v'}]$	$[\overline{v'T'}]$
	($m\ sec^{-1}$)	($m\ sec^{-1}$)	(^{o}C)	$m^2\ sec^{-2}$	$m\ ^{o}C\ sec^{-1}$
			June-July		
100 mb	5.1	4.9	2.8	3.3	0.8
50	3.0	2.1	1.4	0.6	0.4
30	3.2	1.9	1.4	0.1	0.4
20	3.5	2.1	1.5	-0.3	0.4
15	3.6	2.3	1.7	-0.2	0.4
			December-January		
100 mb	7.1	5.7	3.1	10.3	5.2
50	6.7	5.0	2.7	6.1	2.6
30	8.4	5.4	2.9	9.8	3.2
20	9.9	6.3	5.5	18.6	4.0
15	10.2	6.5	3.6	16.2	3.3

associated tropical easterly jet. Since the present study is primarily concerned with the region above 50 mb, further details of the summer circulation of the lower stratosphere will not be discussed herein. An excellent reference for descriptive information on the seasonal circulations at 300, 100, 50, 30 and 10 mb is the series of daily hemispheric analyses for the northern hemisphere presented in Meteorologische Abhandlungen, the publication of the Free University of Berlin.

(b) Winter

The winter circulation is characterized by a deep polar low surrounded by a westerly current. In contrast to the undisturbed zonal flow in the summer easterlies, the westerly flow of the northern hemisphere winter is marked by strong deviations from zonal symmetry. At 100 mb, the disturbances fall into a wide range of wave numbers, but at higher levels the low wave numbers predominate, the smaller scale features having been filtered out. At 10 mb most of the kinetic energy resides in the zonal flow and the lowest two wave numbers, which represent the displacement of the center of the vortex from the pole and the bi-polarity of the vortex.

The long waves of the northern hemisphere stratosphere have certain preferred locations which result in semipermanent features like the Aleutian ridge. Godson and Wilson (1963) have shown that certain preferred regimes tend to recur winter after winter. Hemispheric analyses for these levels bear a strong resemblance to one another from one day to the next and the daily maps often look much like their respective monthly mean maps. In contrast, the daily maps for the troposphere vary markedly over the course of a month, and standing features are recognizable only in the long term averages.

The terms "transient" and "standing" eddies were first applied to planetary scale waves by Priestly (1949) to identify the components of the convariance of two quantities averaged with respect to time and latitude in that order.

$$\underbrace{[\overline{AB}]}_{} = \underbrace{[\bar{A}][\bar{B}]}_{\text{"mean"}} + \underbrace{[\bar{A}^*\bar{B}^*]}_{\text{"standing"}} + \underbrace{[\overline{A'B'}]}_{\text{"transient"}}$$

These terms lend themselves readily to physical interpretation for tropospheric studies in which transient eddies are real entities which are recognizable in the daily maps and vanish in the long term means. However, at high levels in the stratosphere the "transient eddy" term is more a reflection of day to day changes in the position and intensity of the semi-permanent features within the averaging period, and the use of these terms can lead to ambiguities in certain cases. Let us consider, for example, a situation where the stratospheric circulation changes from one preferred regime to another quite different one. If this change occurs in the middle of a month, the transport data are apt to reflect strong transient eddy contributions. The monthly mean map may be relatively featureless due to the cancellation effect and this would result in small standing eddy transports. If, on the other hand, the same change occurred between months, the transient effects would be smaller, and the two monthly maps, in reflecting the features of their respective regimes, would have much more character than in the previous case. This could result in much larger standing eddy transports. Thus it can be seen that the partition of the transports between transient and standing eddies is largely a matter of chance. However, the two terms taken together, still represent the total eddy contribution, and hence the above method of breaking down and evaluating the covariance of two quantities is no less valid than in the troposphere. The only difference is that the distinction between the terms, as suggested by their names, should not be taken too literally.

The data coverage of the southern hemisphere is not sufficient for hemispheric map analyses at these levels and hence the existence of semi-permanent features is still open to question. The lack of standing eddies in the southern hemisphere troposphere (Obasi, 1965) suggests a more uniform zonal flow, less active long waves, and smaller eddy transports than in the northern hemisphere.

Figure 4.1 compares a time height section for the stations in Latitude Band VII (42^{o}N) with a section for Christchurch, N. Z. (43^{o}S). The comparison of a zonally averaged section with one from a single station is justifiable in this case, since the latter section is by far the smoother of the two. If zonal averaging were to effect an unequal smoothing of the two sections, the opposite would be the case. It is apparent from the figure that short term variability is almost absent in the winter circulation of the southern hemisphere, as contrasted with the large month to month variations in the northern hemisphere. At Christchurch the westerlies increase steadily to a maximum in July and decrease smoothly thereafter. If there are disturbances in the polar night vortex they apparently do not extend below 20 mb at middle latitudes. The summer easterlies actually exhibit more irregularity than do the winter westerlies.*

The winter vortex can be divided into two distinct centers of activity in the vertical at middle latitudes:

1. The region below 50 mb, where the westerly flow is the upper most extension of the jet stream. Here the westerly flow and the con-

* This suggests the possibility of a coupling between hemispheres during the northern hemisphere winter. In January of 1961 and '62 when the northern hemisphere winter circulation is interrupted by easterlies there are small minima in the southern hemisphere easterlies. This coincidence could be due to a mean meridional circulation which extends across the equator.

comitant eddy activity both decrease with height.

2. The region above 30 mb which is effectively the lowest extension of the polar night vortex. Here zonal wind speed and eddy activity increase with height.

The intermediate zone is marked by a distinct minimum in the mean westerly flow and in the intensity of the eddy circulations imbedded in it. The zone slopes slightly downward toward higher latitudes and it follows from the thermal wind equation that at any given level it marks the latitude of maximum temperature. Above and poleward of the zone the circulation is thermally direct, with solar radiation supplying energy to the temperature field by differential heating and some of this energy being converted into kinetic energy of the wind field. Below and equatorward of the zone the circulation is thermally indirect, with dynamical processes inducing a temperature gradient which is continually being destroyed by solar radiation. The energetics of this lower region are discussed in detail by Oort (1963), and Newell (1964c) has discussed the energetics of both regions and how they relate to the atmosphere as a whole.

In order to provide some background material on the typical magnitudes of the momentum and heat transports as a function of latitude and level, Fig. 4.2 has been prepared. This shows meridional cross sections of the four year mean values of these parameters. At middle

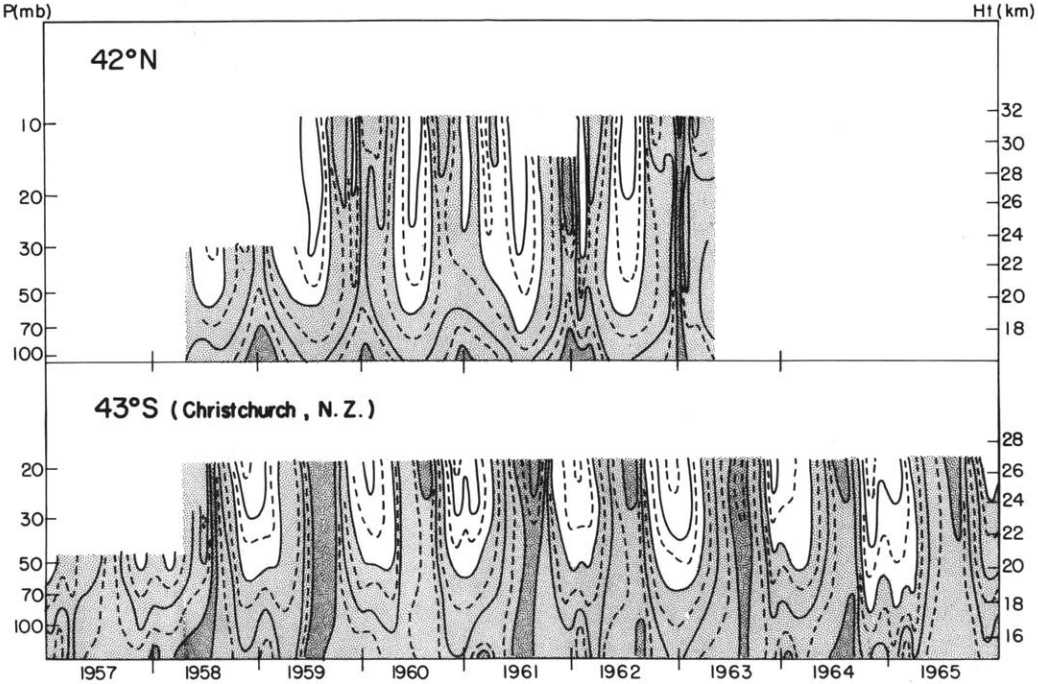
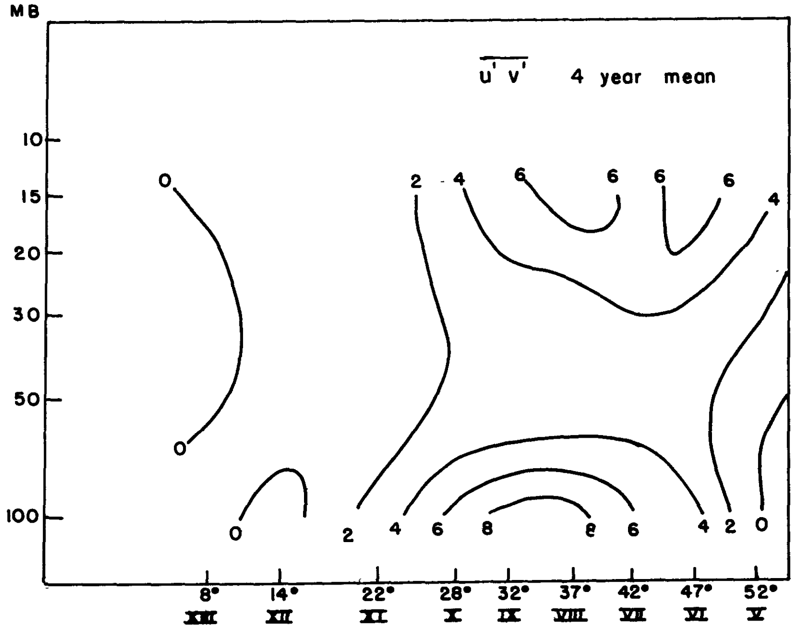


Figure 4.1 Time height sections of zonal wind for latitude band VII (42°N) and Christchurch, N.Z. (43°S). Solid lines are placed at increments of 10 m sec^{-1}. Shaded areas are westerlies. Heavy shading indicates values in excess of 20 m sec^{-1}. December values are placed on tie marks.

latitudes values for the winter months are about double what is on the cross sections while values for the summer months are negligible. The two centers of activity of the winter circulation, and the transition region between them are evident in both sections.

The temporal variations of the winter vortex are often quite spectacular in the northern hemisphere. The zonal wind often undergoes abrupt changes in speed and sometimes the winter westerlies are actually interrupted by periods of easterlies following sudden warmings. (See Fig. 4.1). The heat and momentum flux statistics also show large month to month changes, even to the point of sign reversals. In view of this large short term variability, it would be helpful to examine some parameter at middle latitudes which somehow integrates the effect of the winter circulations over a period of several months, as the zonal wind does in the tropics. It appears that monthly mean total ozone content may be a useful parameter for this purpose.

Newell (1964c) has suggested that the spring maximum in total ozone observed at middle latitudes is due to the increased eddy activity during the winter months, which transports ozone poleward and downward from its photochemical source region. Ozone increases throughout the period of vigorous eddy activity to a peak, the height of which should be a measure of the strength of the winter circulations which produced it. Thus ozone data may provide some indirect evidence as to the relative strength of one winter versus another.



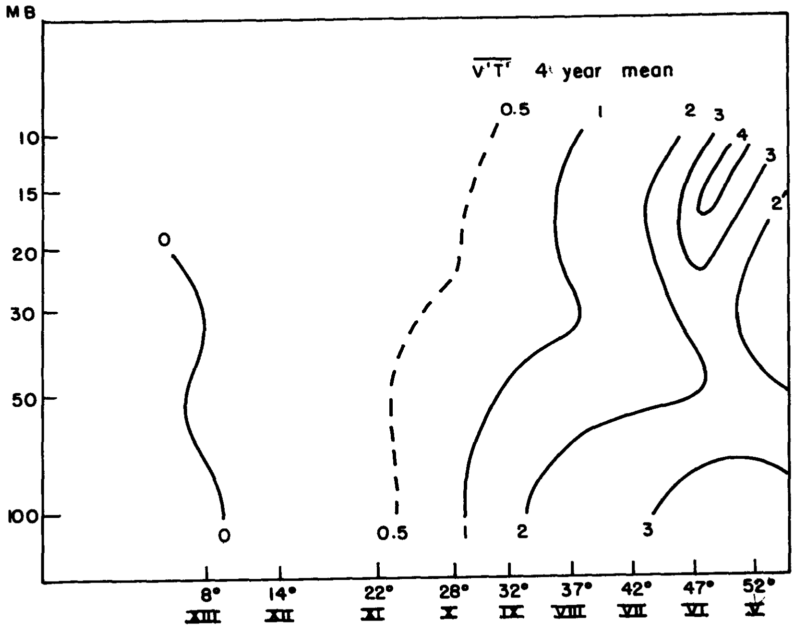


Figure 4.2 Meridional cross sections showing four year means (May 1958 - April 1962) of northward momentum and heat transport. Units in $m^2 \sec^{-2}$ and $m^oC \sec^{-1}$, respectively.

4.2 Long Period Variability

(a) Transient eddy fluxes in the northern hemisphere

Figure 3.5 shows that the long period variations in the momentum transports in the northern hemisphere which give rise to the alternating wind regimes in the tropics extend well into middle latitudes. It will be noticed in Fig. 3.5 that the curves, which represent twelve month running means are somewhat square in shape, with the flat sections centered on the winter months. Now when it is considered how running means are computed it may be seen that this shape can only arise from a situation where all the year to year differences are due to the winter months only. In the previous section it was shown that the momentum transports during the summer season are negligible, and this is apparrently what accounts for the shape of the curves. The long period variations in momentum transports arise then from winter to winter differences and Fig. 3.5 suggests that the transient eddies for the winters of 1958-59, 59-60, 60-61, 61-62 and 62-63 alternate between large and small values.

From Fig. 4.3 it is evident that the year to year variations in momentum transports are accompanied by similar variations in the

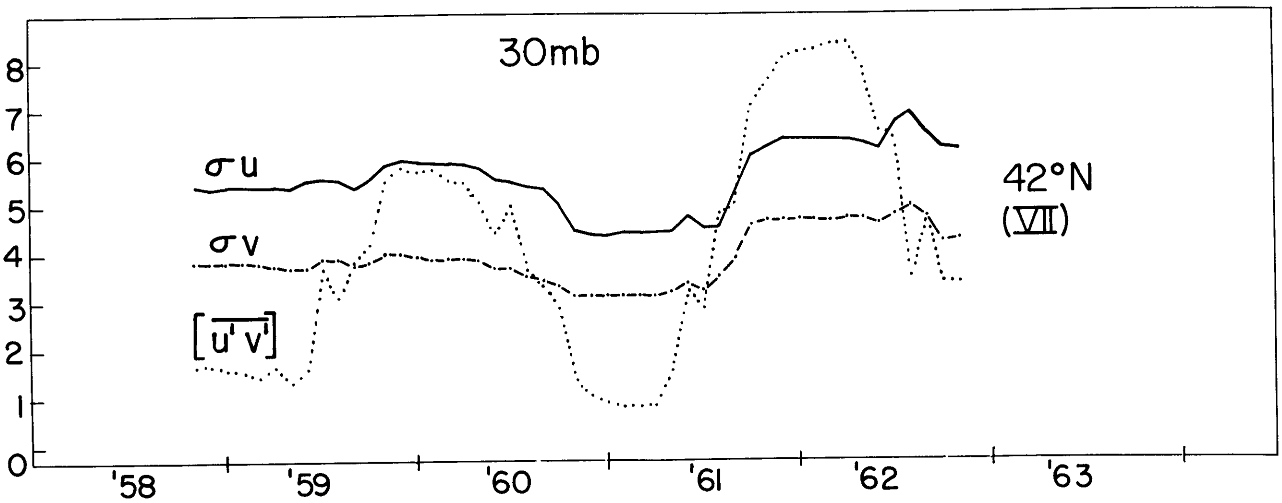


Figure 4.3 Twelve month running means of momentum transport by transient eddies $[\overline{u'v'}]$, together with corresponding curves for the temporal standard deviations of the zonal and meridional components. Units are $m^2\ sec^{-2}$ for $[\overline{u'v'}]$ and $m\ sec^{-1}$ for σu and σv. The running means are centered on the dates indicated. December values are placed on tie marks.

temporal standard deviations of the zonal and meridional wind components. The winters with large poleward momentum transports by transient eddies at middle latitudes are marked by a larger than normal temporal variability of the wind, i. e., more vigorous eddy activity. It should be noted, however, that variations in the amount of eddy activity alone cannot account for all the year to year differences in momentum transport. The subtropics, where the direction of the transport reverses sign from one year to the next give ample proof of the need for variations in the quality of the disturbances (i. e., their shape in the horizontal plane) as well as in the quantity. Tucker (1964) has shown how the wave shape affects the sign of the transport.

Figure 4. 4 shows the unsmoothed monthly values of momentum and heat transport by transient eddies and the temporal standard deviation of the meridional wind component for 37, 42 and 47°N. The tendency for alternating large and small values during successive winters is evident in all three parameters at the higher levels. The winters of 1959-60 and 1961-62 could be classed as "strong" winters with respect to transient eddy phenomena, since they exhibit strong poleward transports of heat and momentum and large temporal variability of the wind. By the same standards, the winters of 1958-59 and 1960-61 would be classed as "weak". The 1962-63 winter is difficult to classify because of the very

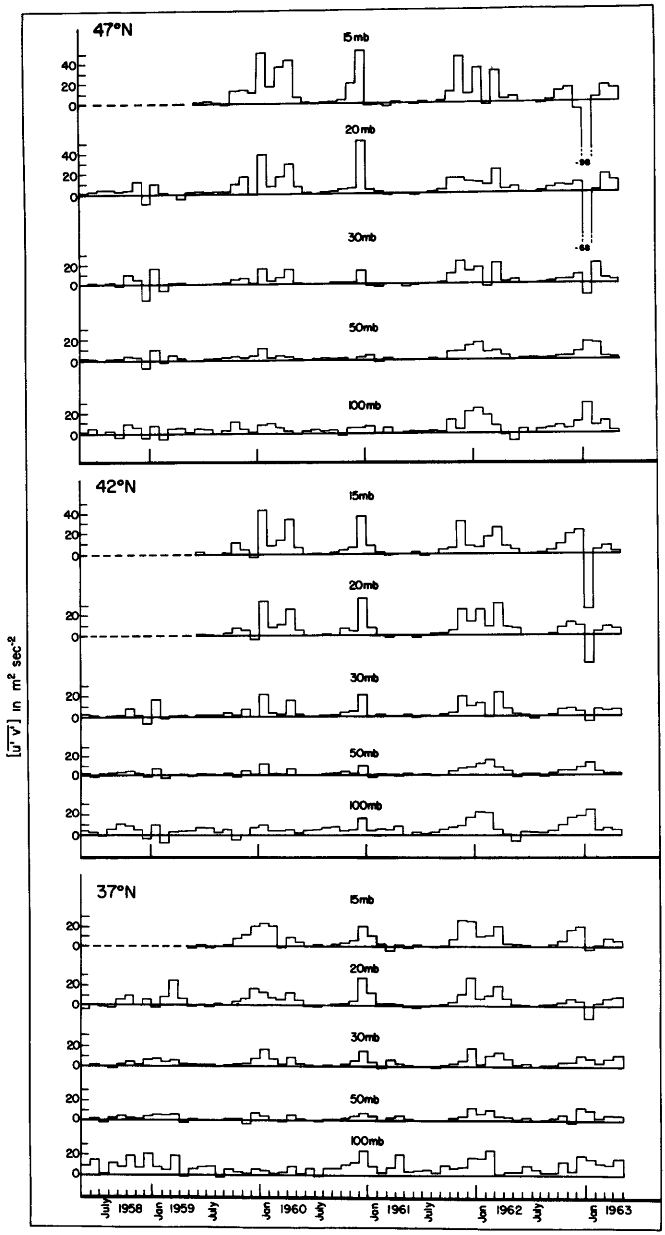


Figure 4.4(a) Monthly values of momentum transport by transient eddies at 47, 42 and 37°N, for several levels. Units are in m^2 sec^{-2}.

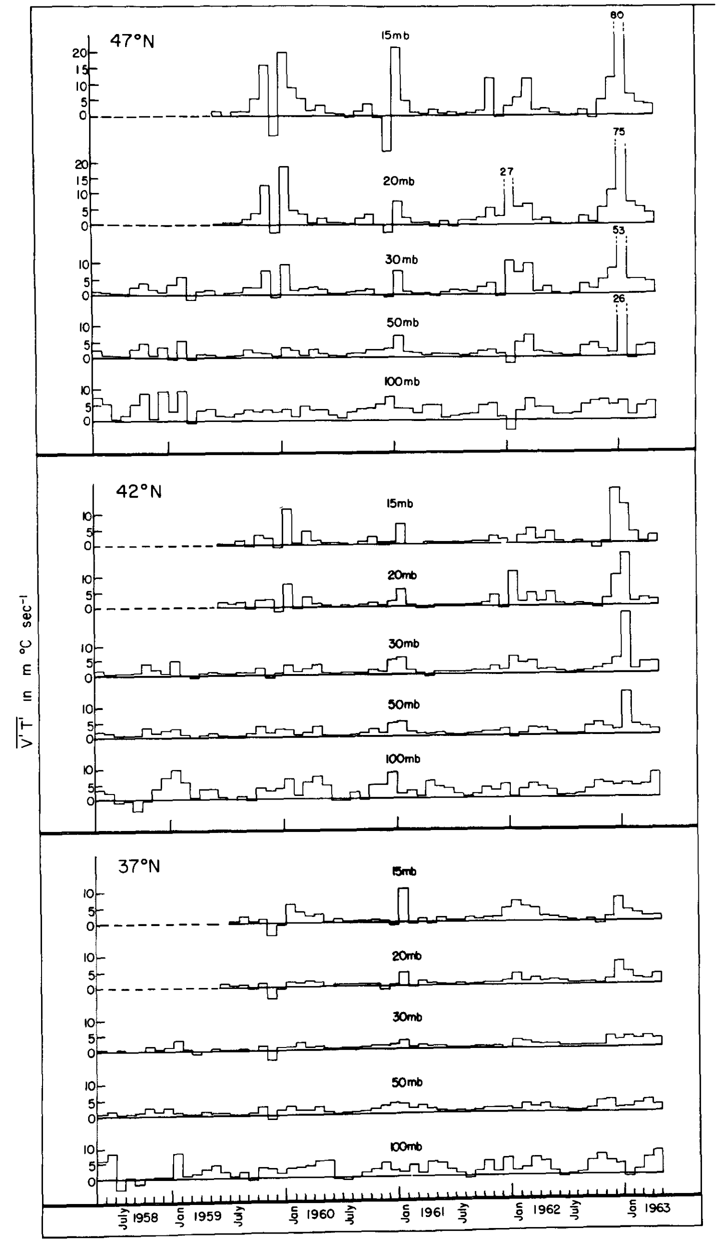


Figure 4.4(b) Monthly values of heat transport by transient eddies at 47, 42 and 37^{o}N for several levels. Units are in m ^{o}C sec^{-1}.

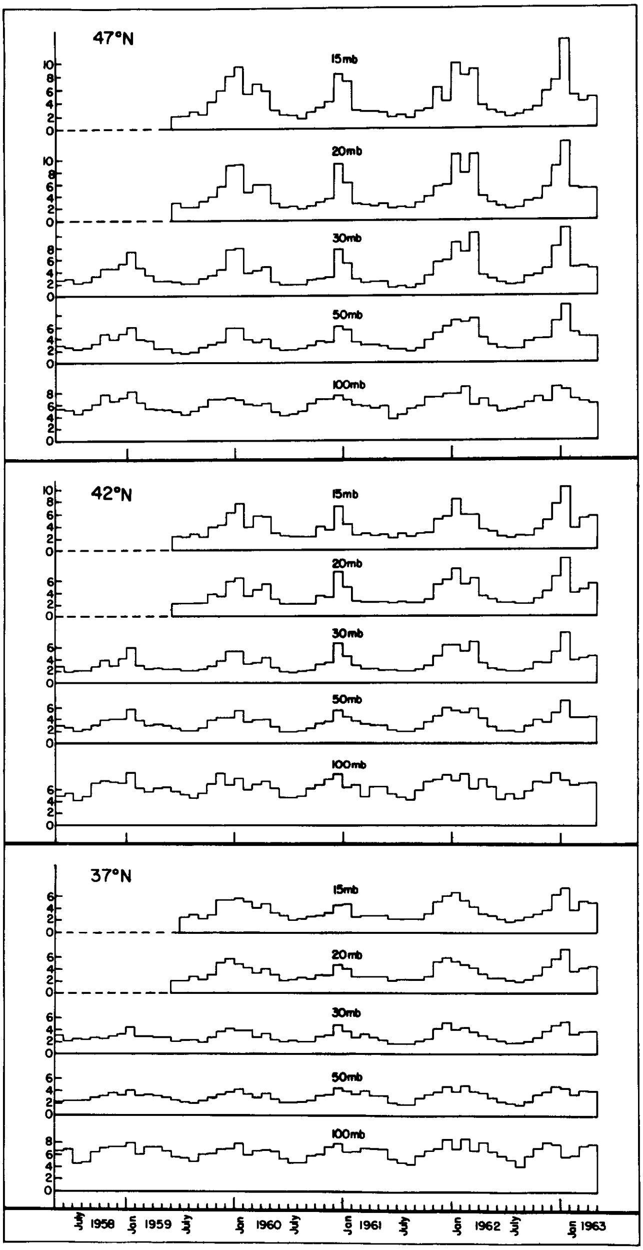


Figure 4.4(c) Monthly values of the temporal standard deviation of the meridional wind component at 47, 42 and 37°N for several levels. Units are in m sec^{-1}.

large values of all three parameters during the month of January which had an unusually strong stratospheric warming.

The reliability of the transient eddy statistics is somewhat open to question, because the distribution of data is highly asymmetric with respect to longitude. (See Fig. A). The complete lack of high level observations during this period from Russia and China, coupled with the sparsity of data over the underdeveloped countries and ocean areas creates a situation where most of the data are concentrated in three regions: North America, Europe and Japan. There is no way of knowing for certain whether the statistics as computed, reflect the time variations in the true zonally averaged quantities. In their defense all that can be said is that the statistics, as computed, display a pattern which is consistent with respect to latitude and height, and, as will be discussed in the Sec. 4.3, fit well with the momentum, heat, and ozone budgets of the region.

(b) Standing eddy fluxes in the northern hemisphere

It was possible in the tropics to demonstrate that standing eddies play only a minor role in transport processes as compared with transient eddies. However it is clear from the discussion in Sec. 4.1b that at middle latitudes, at least in the northern hemisphere, the effects of standing eddies on transport processes cannot be ignored. Computations of seasonal wind and temperature statistics by the MIT Planetary Circulations Project, based on IGY data (for a summary, see Oort, 1963) show that at middle and high latitudes, the transient and standing eddy effects are of comparable magnitude. It is to be expected that the shorter averaging period used in the present study would serve to increase the relative importance of the standing eddies.

The monthly mean hemispheric analyses of the geopotential height field published in Meteorologische Abhandlungen were used as a basis for computing the spatial standard deviation of the monthly mean meridional wind component taken around a latitude circle, $\sqrt{[\bar{v}^{*2}]}$, and the momentum transport by standing eddies $[\bar{u}^* \bar{v}^*]$. The procedure used for calculating these quantities is discussed in the appendix. Unfortunately, at the time of this writing, monthly mean map analyses were not yet completed for the period 1960-63. Therefore the only period available for comparison with transient eddy data is May '58 - Dec '59. Table 4.2 shows a comparison of the transient and standing eddy statistics for the months of the 1958-59 winter. (The levels and latitudes for which the eddy statistics were computed are not compatable, so it was necessary to compare 47, 37 and 28°N for the transient eddies with 45, 35 and 25°N, respectively for the standing eddies, and 30 mb for the transient eddies with 25 mb for the standing eddies).

The statistics show that transient and standing eddy effects are of

Table 4.2. Comparison of transient and standing eddy statistics for the 1958 - 59 winter.

	30 mb				50 mb			
	Standing		Transient		Standing		Transient	
	σ_v	$[\bar{u}^*\bar{v}^*]$	σ_v	$[u'v']$	σ_v	$[\bar{u}^*\bar{v}^*]$	σ_v	$[u'v']$
45°N								
Oct	4.3	5.4	4.5	10.2	4.1	5.7	4.6	4.3
Nov	7.8	19.2	4.5	4.9	6.6	22.4	3.7	2.9
Dec	10.9	14.3	5.4	-15.2	9.6	-3.0	4.7	-8.8
Jan	12.6	-22.5	7.4	16.5	12.1	-40.9	5.8	9.7
Feb	10.8	15.3	4.7	-6.0	9.0	2.3	3.9	-2.4
Mar	8.6	25.4	3.4	1.4	6.8	9.2	3.6	4.7
Apr	3.2	4.1	2.5	1.8	4.6	0.0	2.8	1.6
35°N								
Oct	2.8	5.1	2.5	4.1	2.9	3.6	3.1	2.2
Nov	3.9	8.5	2.9	0.5	3.9	14.5	3.6	1.5
Dec	5.8	14.6	3.2	6.0	4.6	10.7	3.2	3.3
Jan	6.6	-5.1	4.3	7.2	5.6	-11.7	4.0	4.6
Feb	5.0	11.6	2.9	3.4	4.2	3.4	3.1	4.0
Mar	4.5	14.7	2.8	5.3	3.4	6.0	3.4	5.9
Apr	1.9	1.9	2.7	1.6	2.8	-0.3	3.1	-2.7
25°N								
Oct	2.2	1.1	2.7	2.0	2.2	-0.2	3.0	2.3
Nov	3.0	1.4	2.6	1.9	3.5	4.4	2.8	3.7
Dec	4.2	-2.2	2.8	-0.6	3.7	7.9	3.2	10.1
Jan	4.2	-1.6	3.0	7.1	2.0	-1.3	3.5	2.3
Feb	3.9	4.8	3.6	7.5	2.5	0.7	3.5	6.1
Mar	3.2	4.3	2.4	0.9	2.7	1.4	3.4	2.8
Apr	1.8	1.4	3.1	1.6	2.0	-0.1	3.4	-2.6

comparable magnitude at 25°N, but the relative importance of the standing eddies increases rapidly with latitude. This suggests that at high latitudes the transient eddy statistics are not a suitable index of the general level of eddy activity. For this reason, data on transient eddies are not presented for latitudes poleward of 47°N. However, at lower latitudes, where the transient eddies contribute a substantial fraction of the total, it is likely that the winter to winter variations in their intensity reflect a real variability in the strength of the eddies as a whole. If this were not the case, there would have to be some systematic negative correlation between the monthly values of the standing and transient eddies. There is no physical reason for expecting large poleward fluxes by standing eddies to accompany small poleward (or equatorward) fluxes by transient eddies, and the data in Table 5.1 indicate no such tendency. On this basis, it is assumed in the following discussion that the transient eddy statistics are adequate for describing qualitatively the winter to winter variations in the overall level of eddy activity.

It will be interesting to compare the transient and standing eddy statistics over a longer time interval and to see whether the standing eddies also exhibit winter to winter variations. It is hoped that monthly mean maps for the missing years will soon be published. As a word of caution, it should be noted that the standing eddy momentum fluxes, as computed from monthly maps are subject to some serious uncertainties because of sparse data coverage. Momentum transports depend upon the tilt of the waves which is, in regions of sparse data, often a matter of the analyst's artistic taste. $\sqrt{[\overline{v^{*2}}]}$ should prove a more reliable index of long period variations.

(c) Eddy fluxes in the southern hemisphere

There is to date no direct evidence of winter to winter differences in the eddy fluxes in the southern hemisphere. However, long term variations in ozone and temperatures at middle latitudes, which will be discussed in Sec. 6.3, suggest an alternating series of weak and strong winters as is observed in the northern hemisphere.

(d) The heat, momentum and ozone budgets

It was shown in the previous chapter that the behavior of the temperature field in the tropics is probably governed, for the most part, by mean meridional circulations, and eddy effects are relatively small. However at middle latitudes, where the eddy heat fluxes and their divergences are about an order of magnitude larger than in the tropics it is likely that these disturbances exert a noticeable influence on the temperature distribution. It can be seen from Fig. 5.1 that the warm winters at middle latitudes of the northern hemisphere are those winters which have exhibited strong eddy activity. This is consistent with the notion that poleward heat transport by the eddies is responsible for the warm temperatures at middle latitudes in the winter stratosphere.

It will be demonstrated in the next chapter that the wind at high levels of the northern hemisphere during the winter seasons exhibits a behavior which can best be described as an effect of the variations in the transport of momentum by the eddy circulations.

For reasons described in Sec. 4.1b the peak of the spring maximum in ozone concentration is an index of the strength of the winter circulation. Ramanathan (1963) produced evidence of a two year periodicity in the spring ozone maxima at middle latitude stations of both hemispheres. The data are reproduced in Fig. 4.5. It will be noticed that the "strong" winters of 1959-60 and 1961-62 are marked by relatively high ozone concentrations in the northern hemisphere. It is also apparent from comparing Figs. 4.5 and 5.1 that the warm southern hemisphere winters are those with high ozone peaks. This is also suggestive of a response to variable eddy intensity, with the winters of even years being the strong ones. Ramanathan's data suggest that this tendency for large ozone winters to alternate with small ones has prevailed throughout much of the past decade.

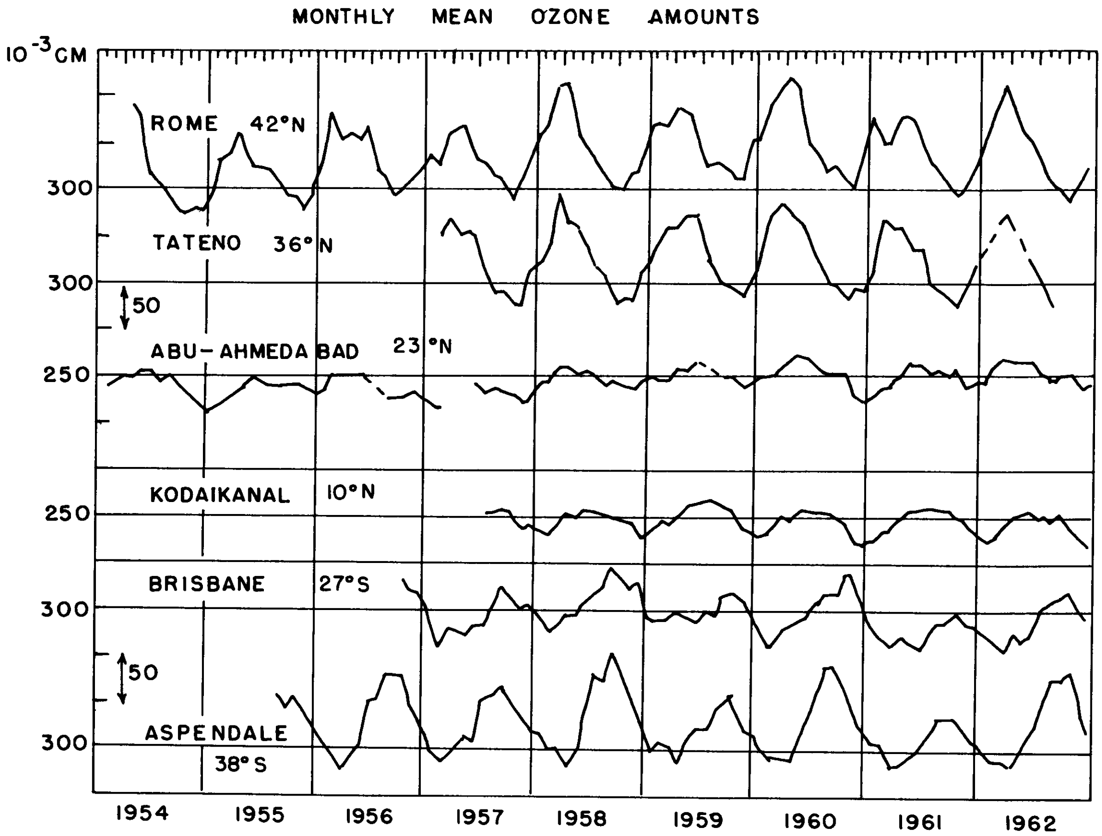


Figure 4.5 (After Ramanathan (1963)).

4.3 The Two Year Periodicity

(a) The nature of the phenomenon

It is interesting to note that throughout the period 1957-62 when, as noted in the previous chapter, the wind regimes in the tropics repeated themselves at two year intervals, the winter circulations also exhibited a two year periodicity in both hemispheres. This was evidently an interval during which the long period variations were phase locked with the annual cycle. The long record of ozone data from Arosa (Fig. 4.6) shows that this has not always been the case. Prior to 1953 the ozone peaks fail to show any two year periodicity, and since 1963 there appears to be a breakdown of the periodic pattern of the previous decade. All middle latitude stations in both hemispheres which exhibited the periodicity during the 1953-63 decade shared the subsequent breakdown, and the winds in the tropics began to depart strongly from their periodic behavior beginning in 1963. This suggests that 1963 marks the end of an interval of phase locking with the annual cycle and the beginning of a period when the variations of the winds in the tropics are somewhat more irregular as they respond to what appears to be a series of winters whose intensities do not follow any regular pattern .

(b) The question of statistical significance

The question arises as to whether this decade of alternating

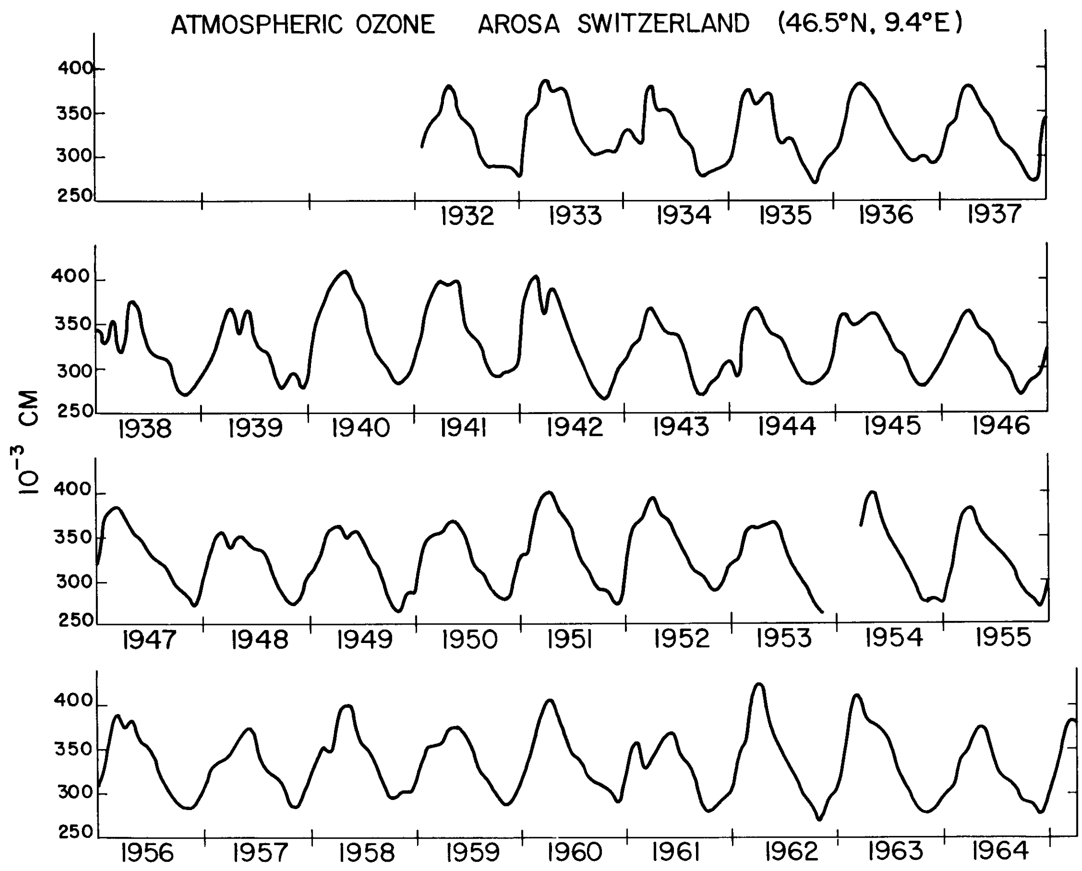


Figure 4. 6 Mean monthly ozone amounts at Arosa, Switzerland 1932-1965. Units in 10^{-3} cm @ STP.

high and low ozone years and the related five year series of alternating "strong" and "weak" winters represents a statistically significant tendency for a two year oscillation, phase locked with the annual cycle, or whether it is possibly a chance occurrence which requires no physical explanation. Unfortunately, the data for Arosa (Figure 4. 6 and Table 4. 3) represent the only unbroken ozone record which extends into the period prior to 1953 when the decade of alternating winters began. There is some question as to whether these data are truly representative of the earlier period. Even within the 1953-63 decade, the periodicity in the Arosa data is not as impressive as that from other stations (see Figure 4. 5). Table 4. 3 shows that the sequence of alternating values in the Arosa data was broken by the 1956 winter, while Rome and Aspendale show no such lapse. Moreover, the reliability of the earlier data may be questioned, since current measurement techniques were not employed prior to the 1950's. Thus it is perhaps too early to make any statement regarding the statistical significance of the two year periodicity at stratospheric levels. However, since tropospheric data suggest that such a periodicity is present in the atmosphere (see Ch. 5) it might be well to ask how such a phenomenon might be caused.

(c) Possible causes

The atmosphere contains within itself a number of mechanisms which may be capable of producing the observed tendency for a two year oscillation. The relatively undisturbed zonal current in the tropics ex-

Table 4. 3. Largest monthly mean value of total ozone in each year.

Units: cm x 10^{-3} at S. T. P.

Arosa, Switzerland (46^{O}N - 9^{O}E)

	0	1	2	3	4	5	6	7	8	9
192_								367	360	401
193_			378	384	377	374	382	379	377	370
194_	412	401	405	370	370	363	366	382	356	363
195_	367	400	394	367	400	384	381	367	393	376
196_	406	371								

periences change on a much longer time scale than the rest of the atmosphere because friction is so small. Consequently the zonal flow at these levels "remembers" events which occurred a year previous. As an example, let us suppose that a "strong" winter with large poleward momentum fluxes gives rise to an easterly wind regime at the 10 mb level during a given winter. If the descent rates as deduced from Fig. 3.1 (at 3^{o}S) are typical, the following winter the same regime will be found in the neighborhood of the 50 mb level. In other words, the flow at 50 mb in the tropics will be a reflection of the strong winter which occurred a year previously. Now if there is any mechanism by which the zonal flow in the tropics can exert a negative feedback on the eddy circulations, so that, for instance, the easterly regime at 50 mb tends to favor a weak winter at middle latitudes - the type of winter opposite from that which produced it - this is all that is required for a two year cycle. In the hypothetical case in point, the weak winter favored by the easterlies at 50 mb would produce westerlies at 10 mb. By the next winter these would descend to 50 mb, and their location there would tend to favor a strong winter, which would produce easterlies at 10 mb, and so on.

Now let us consider the various means by which the mean zonal flow might produce such a feedback effect upon the eddy circulations of the winter season.

The direction of the zonal flow in the tropics has a noticeable effect on the appearance of the entire hemispheric circulation patterns during the winter season. Figure 4.7 contrasts the months of March 1958 and 59. In the former year the flow at low latitudes is westerly. A belt of high pressure covers the tropics and the polar vortex occupies most of the hemisphere, with westerly flow poleward of 20°N. In the latter year easterly flow in the tropics displaces the high pressure belt northward into the subtropics, which is consistent with geostrophic considerations. This confines the polar vortex to narrower limits than in the previous year, and westerly flow is found only poleward of 30°N. These year to year differences are typical of all the months in the 1958 and 1959 winters.

This southward limit of the winter westerlies may affect the transmission of eddy energy from the troposphere into the stratosphere. Charney and Drazin's (1961) theoretical results indicate that westerly flow is essential for the upward propagation of the energy of planetary waves. If these results are truly applicable to this situation, and if the eddy disturbances in the upper, thermally direct region of the stratosphere receive a significant amount of their energy from the troposphere, then the effect of the zonal wind profile upon the upward transmission of wave energy could play an essential role in producing the observed two

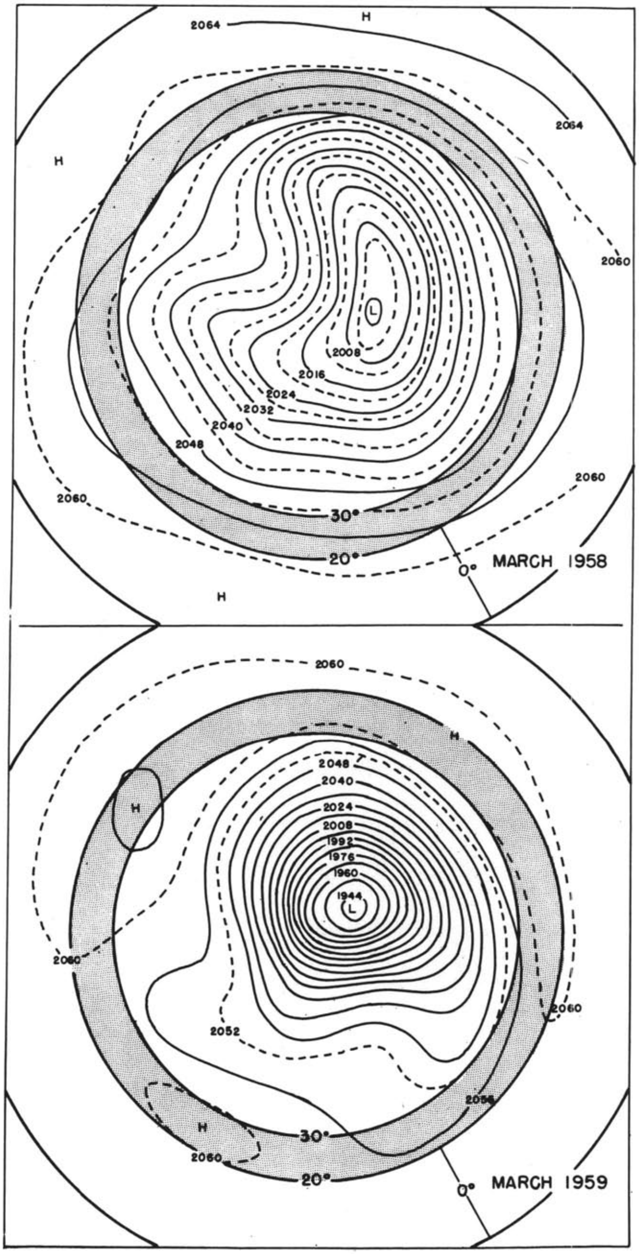


Figure 4.7 Monthly mean geopotential height analyses of the 50 mb surface for March 1958 and March 1959, as adapted from Meteorologische Abhandlungen.

year periodicity.

To test this hypothesis, monthly values of the correlation coefficient between the meridional wind component at 50 and 100 mb were computed for a number of latitude bands. It was found that at 22, 28 and 32°N these showed a consistent year to year variation, with high values during the 1959-60, and 1961-62 winters and low values during the 1958-59 and 1960-61 winters (Fig. 4.8). It can be seen from the section for 20°N in Fig. 3.1 that in the layer between 50 and 100 mb the former winters are marked by westerlies and the latter ones by easterlies. Thus these results are in agreement with Charney and Drazin's findings. Although the year to year variations in the interlevel correlations are not large and they do not extend to levels above 50 mb, the fact that they are in the right sense at three independent latitudes is encouraging.

This then is one possible explanation for the tendency for a two year periodicity in the stratospheric circulation. The essential requirements for it to work are (1) the mean downward motion in the tropics which transports wind regimes from 10 mb to around 50 mb over the course of a year, (2) the ability of the mean zonal flow to regulate the upward transport of wave energy from the troposphere and (3) the dependence of the upper level disturbances upon energy from the troposphere.

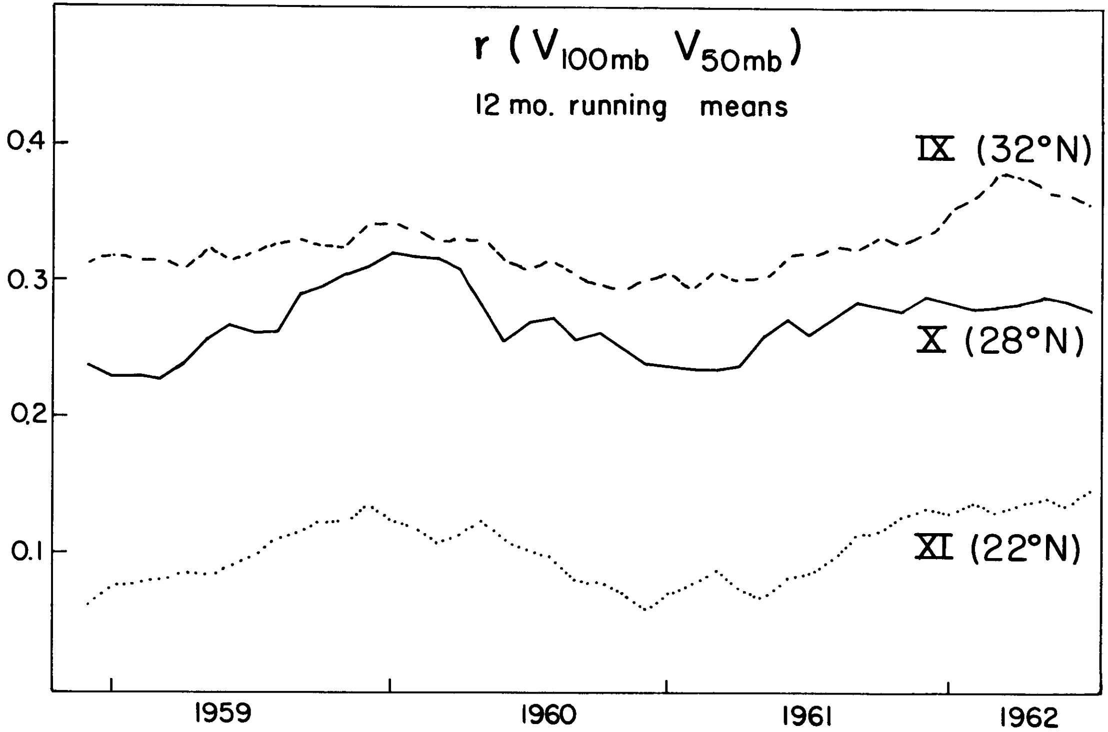


Figure 4. 8 Twelve month running means of the zonally averaged interlevel correlation coefficient between the meridional wind components at 50 and 100 mb at three latitudes. Units are dimensionless. Running means are centered on dates indicated. December values are placed on tie marks.

It is possible to conceive of other mechanisms by which the mean zonal flow could influence the eddy circulations. As an example, it was shown in the previous chapter that the momentum fluxes are usually up the gradient of zonal wind. This condition usually arises wherever the flow is roughly two dimensional and the eddies are on the same (space) scale as the gradient of the zonal wind. Indeed, it is simple to demonstrate that the shear in the mean flow will always distort such eddies so as to cause them to transport momentum up the gradient. Thus it is possible for the mean zonal flow, as it changes in response to the momentum fluxes to exert a feedback upon the eddy disturbances, which influences their shape. This effect may be responsible for the alternating direction of the momentum transport at low latitudes.

The mean zonal flow at low latitudes may also influence the eddy disturbances in that it determines, to a large extent, the boundary condition at the outer edge of the polar vortex. It is possible that the stability criteria governing the growth rate of disturbances within the vortex are sensitive to this boundary condition, in which case, this could prove to be a very important mode of interaction.

The above discussion of feedback mechanisms has been confined to the stratosphere of one hemisphere for purposes of simplicity. The questions of interactions between hemispheres and between troposphere and stratosphere will be reserved for the concluding chapter.

4.4 The Modeling Problem

The relationships between the eddy fluxes and the budget of heat, momentum and ozone at various latitudes as discussed in this chapter are shown schematically in Fig. 4.9. The dashed line connecting the mean zonal flow in the tropics with the eddy fluxes represents the suggested feedback effect.

Evidence has been brought forth to show that there is a tendency for winters in the stratosphere to alternate between strong and weak eddy circulations. The investigation into the possible causes of this phenomenon suggests that the atmosphere may, of itself, contain all the necessary ingredients for producing a phase locking with the annual cycle:

1. A strong annual cycle in the stratosphere with a short very active winter season and a quiet summer season.

2. The capacity for "remembering" from one active winter season to the next. The "memory" is contained in the mean zonal circulations of the tropics.

3. A means for the active winter circulations to interact with the memory, so as to store information on the characteristics of one winter

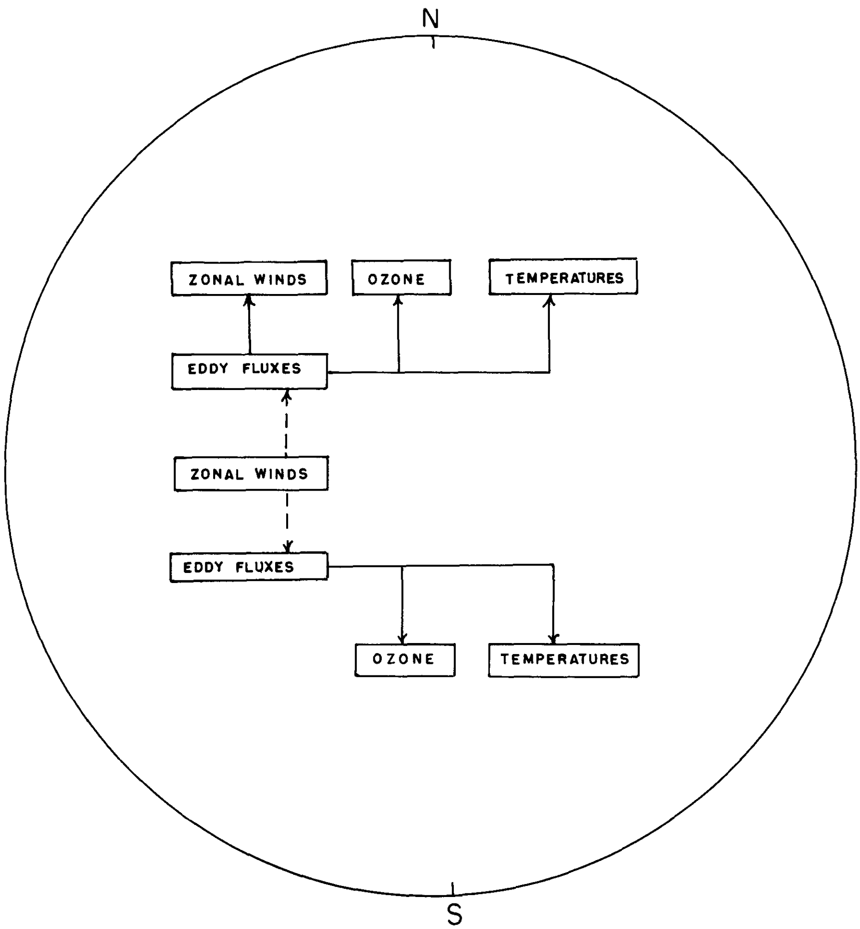

N
ZONAL WINDS
OZONE
TEMPERATURES
EDDY FLUXES
ZONAL WINDS
EDDY FLUXES
OZONE
TEMPERATURES
S

Figure 4.9

(i.e., through the momentum transports) and to react to this information the next winter.

4. The possibility for this interaction with the memory to produce a negative feedback so that a vigorous winter one year will tend to produce a quiet one the next and vice versa.

There are a number of possible mechanisms by which the atmosphere may transmit information on the previous winter season from the memory back into the winter circulation. This is the least known link in the chain of cause and effect and undoubtedly the most difficult on to model accurately. It would be impossible to parameterize the eddy circulations in a model of the two year oscillation without making some assumption as to how this feedback mechanism works, and to make such an assumption would be to defeat the very purpose of such a model. A more informative experiment would be to allow the eddy disturbances to develop spontaneously in an initial value problem which could be integrated over a period of many years. Perhaps a simplified model containing only a few zonal wave numbers such as that of Peng (1965), coupled with a model such as that described in Sec. 3.7 for dealing with the mean zonal and meridional circulations would contain all the necessary ingredients for simulating phase locking with the annual cycle.

CHAPTER V

LONG PERIOD VARIATIONS IN OTHER PARAMETERS

There are certain parameters not discussed in previous chapters which also exhibit long period variations. It is difficult to relate these parameters directly to the dynamical processes considered in Chapters III and IV, and therefore to discuss them will contribute only in a minor way to our understanding of the underlying mechanism of the "biennial oscillation". Nevertheless, it is worthwhile doing so, for the sake of completeness, and to point out certain relationships within the atmosphere which deserve further study.

The subject of two year or 26 month periodicities has received a great deal of attention in the recent literature. Time records of literally dozens of parameters have been scrutinized, usually by means of spectral analysis techniques, with hopes of finding peaks near 26 months. Because the choice of quantities to be investigated was often made on the basis of what long period records happened to be available and without recourse to physical reasoning, the list of variables presently alleged to show 26 month periodicities includes such diverse quantities as tree ring thicknesses (Bryson and Dutton (1961)), lake levels (Wallen (1913)), cosmic ray intensity (Maeda (1966)), solar diameter (Newell (1964a)), stratospheric warmings (Labitzke (1966)), and hurricane frequency (Hanna (1965)) a collection which defies any unified physical interpretation. All that can be safely said is that these studies have yielded positive results in a sufficiently large number of cases that it is difficult to ascribe the widespread appearance of such periodicities to chance.

It is important to distinguish between those quantities which exhibit periodicities which are apparent only in spectral analysis of long data records, spanning several decades or more, and those which, over the past decade have shown a strong persistent correlation with the zonal wind in the tropics. While the former quantities may show real periodicities, it is by no means certain that they are related to the tropical wind oscillation. Even if such a relation exists it is likely to be a weak one. The following discussion will be concerned primarily with those quantities which show evidence of variations associated with those in the tropical stratosphere. It is convenient to divide the parameters to be considered into two groups, depending upon whether they are chiefly observable in the stratosphere or troposphere.

5.1 Parameters in the Middle and High Latitude Stratosphere

(a) Temperature

Angell and Korshover (1962, 63) were the first to discuss the extension of the "biennial oscillation" into middle and high latitudes at stratospheric levels. Their pole to pole representation of twelve month

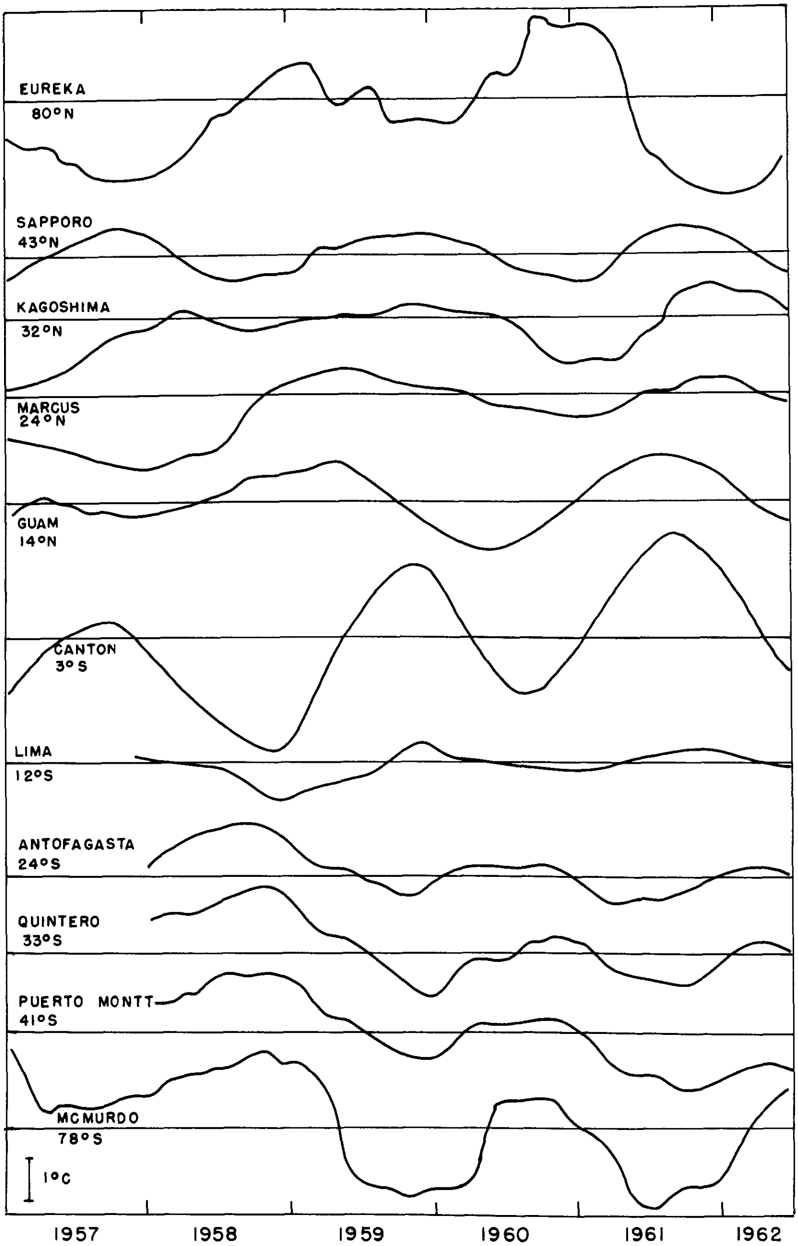


Figure 5.1 Twelve month running means of 50 mb temperatures centered on dates indicated. (After Angell and Korshover (1964)).

running means of temperature at 50 mb (Fig. 5.1) is one of the most fascinating and perplexing pieces of evidence available on the subject. From their analysis, the following points are apparent:

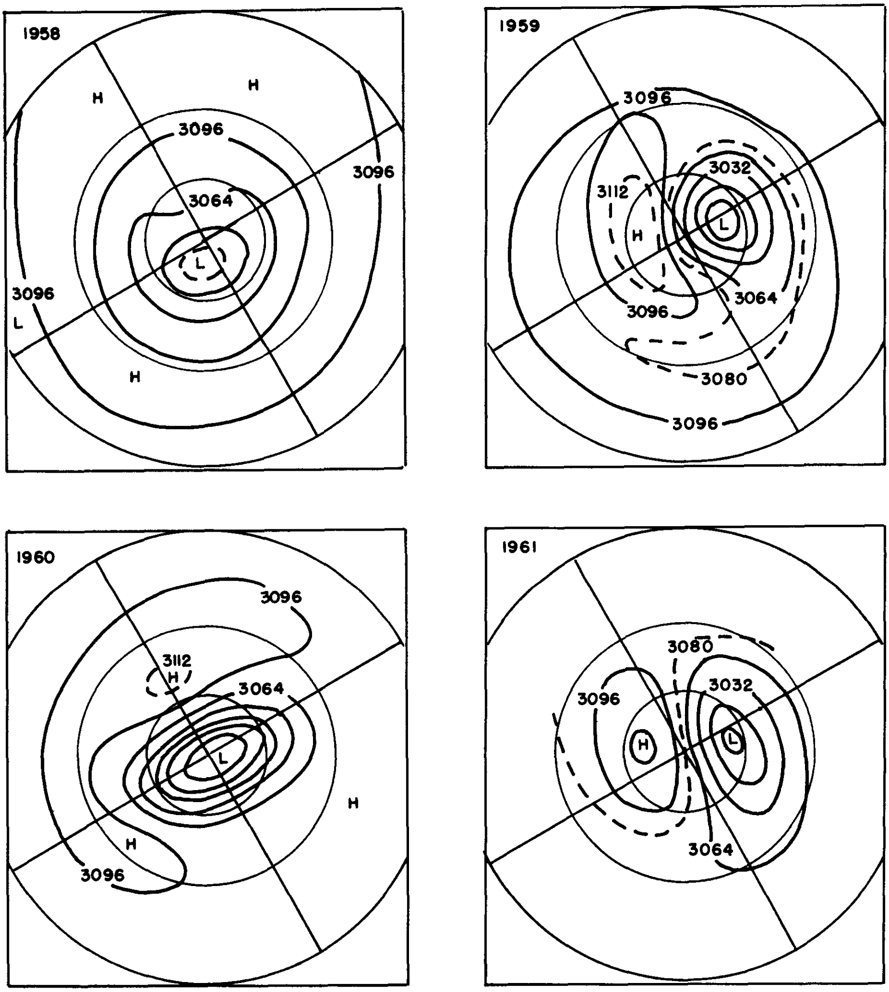
1. At middle latitudes of the northern hemisphere the temperature maxima occur near the beginning of the even years, about six months earlier than at the same latitudes in the southern hemisphere. In both hemispheres the temperature maxima are approximately concurrent with the ozone maxima at these latitudes (see Fig. 4.5) which is to be expected if the year to year variations are caused by enhanced poleward fluxes of ozone and heat during alternate winters, as suggested in Sec. 4.2d.

2. In the northern hemisphere there is a phase reversal between middle and high latitudes; i.e., the polar stratosphere is cold during the winters that middle latitudes are warm, while in the southern hemisphere there is no such reversal.

3. The 1957 winters at the poles do not appear to conform to the two year periodicity.

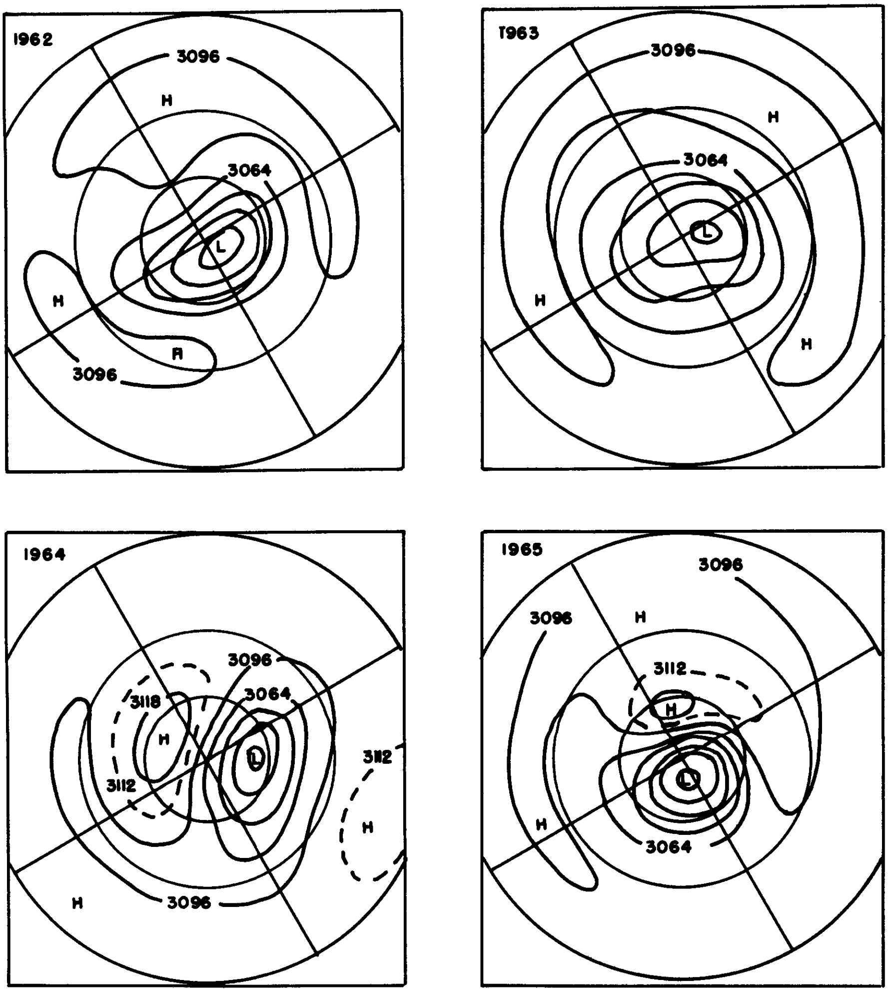
(b) Final warmings

The behavior of the high latitude temperatures becomes more plausible when one considers more recent findings by Labitzke (1966) and Barbe and Reininger (1966) on the nature of the spring warmings in the two hemispheres. Labitzke contends that there is a strong relationship between the phase of the wind oscillation in the tropics and the timing



Monthly mean maps for March
1958 - 1961

Figure 5. 2(a) (After Labitzke (1966))



Monthly mean maps for March
1962 - 1965

Figure 5. 2(b) (After Labitzke (1966))

and nature of the final warmings in the northern hemisphere stratosphere. As evidence she cites the 10 mb northern hemisphere maps for the months of March for eight successive years (Fig. 5.2). The difference between the years 1958, 1960, 62, 63, 65 and 1959, 61, 63, 64 is readily apparent. In the first set of years the polar vortex remains centered over the Arctic and the hemispheric circulation retains a considerable amount of zonal symmetry, while in the second set the vortex is displaced toward Siberia. Wave number one is important in the profiles of both sets of years but, as Labitzke points out, its amplitude is more than twice as large in the latter set, at the higher latitudes. There is also a slightly different phase of wave number one in the two sets of years; in the former the ridge position is over the Aleutians while in the latter it is positioned over western Canada.

The data available for the southern hemisphere warmings is much more limited. According to Barbé and Reininger (1966), the 1963 warming in the Antarctic occurred during the middle of November, a month later than the 1964 warming. There are no stratospheric analyses available prior to 1963; in the absence of such data, the best indicator of the time of the stratospheric warmings is the 100 mb temperature at Antarctic stations. Phillpot (1964) has presented the 100 mb temperature records at a number of Antarctic stations for the years 1957-62; two of these are

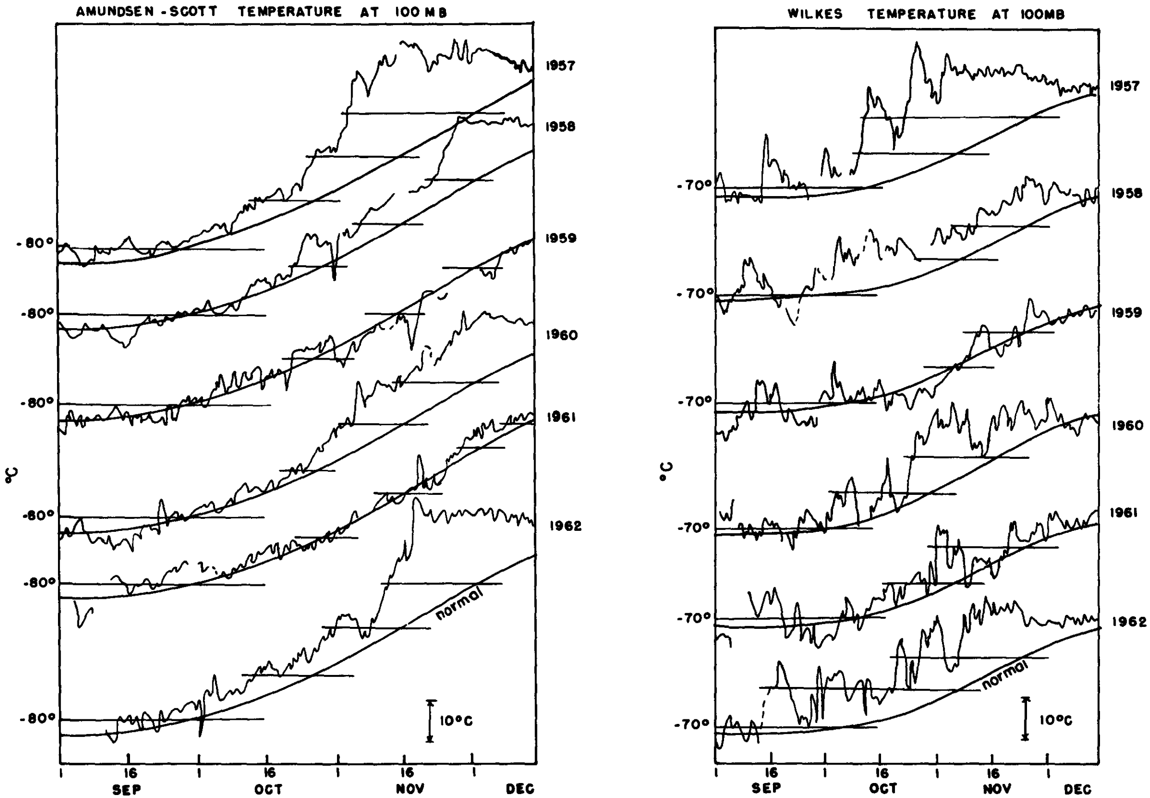


Figure 5. 3 100 mb temperatures at Amundsen-Scott and Wilkes during spring warmings, 1957-1962.

Units are in °C. (After Phillpot (1964)) shown in Fig. 5.3. These suggest that the years 1957, 58, 60 and 62 were marked by relatively early or "accelerated" warmings, as opposed to 1959 and 61 in which the warmings occured about a month later. Phillpot's 50 mb data from Amundsen-Scott also suggest the same behavior.

The reason for the oscillation in the polar temperatures now becomes quite apparent. In winters with early warmings the stratospheric temperatures at high latitudes rise to summer levels about a month earlier than in the other winters. This early rise is reflected in the twelve month running means as a slightly increased temperature for the entire year centered around the month of the warming. In all cases where data on both polar temperatures and the time of the final warming are available, the years with warm winters are found to be those with early warmings. These results are summarized in Table 5.1

(c) Zonal winds

The time variations of zonal wind at middle latitudes in the stratosphere also show strong evidence of a two year periodicity. Rofe (1963) and Sparrow and Unthank (1964) have published data for Australia and New Zealand, some of which has been reproduced in Fig. 5.4. The oscillation is quite pronounced at 100 mb, and the square wave pattern centered on the winter months suggests that most of the year to year variations are due to the winter seasons.

Table 5.1 Summary of stratospheric winters 1957-64.

Year and hemisphere		Time of final warming	Polar temp 50 or 100 mb	Mid lat. zonal wind	Shear parameter in tropics
1957	N		medium		
1957	S		warm	W	-
1958	N	late	cold		-
1958	S		warm	E	+
1959	N	early	warm	E	+
1959	S		cold	W	-
1960	N	late	cold	W	-
1960	S		warm	E	+
1961	N	early	warm	E	+
1961	S		cold	W	-
1962	N	late	cold	W	-
1962	S		warm	E	+
1963	N	late		W	+
1963	S	late	cold	W	0
1964	N	early			-
1964	S	early	warm		

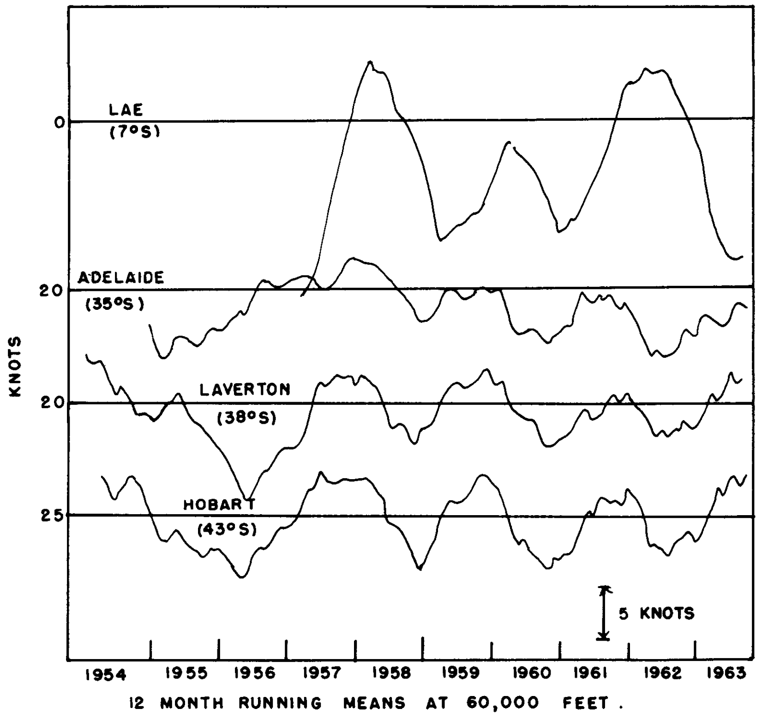


Figure 5.4 Twelve month running means of zonal wind at 60,000 ft. for Australian stations. Units are in knots. Running means are centered on the dates as indicated. (After Sparrow and Unthank (1964))

Angell and Korshover's (1963) analysis of northern hemispheric winds failed to produce such clear cut results. There was some evidence of a periodicity at some stations but the amplitude was very small. A more detailed examination of this region was included in the present study with hopes of obtaining more definitive information on the nature of the year to year variations. Monthly means at individual stations were averaged together into latitude bands, as shown in Fig. 2.1. (See appendix for details). Twelve month running means were computed from the averages for the latitude bands and these are shown in Fig. 5.5. A consistent pattern of year to year variations appears over a wide range of latitudes and levels in this figure. Winter to winter differences account for most of the variance of the curves, as in the southern hemisphere. However, unlike the southern hemisphere, the oscillation is noticeable only above 30 mb at the higher latitudes.

Barbé and Reininger (1966) have suggested that the year to year variations in the southern hemisphere winds are only a manifestation of the difference in the time of the final warming from one year to the next. Years with early warmings are marked by a premature end of the winter westerlies and an early onset of the summer easterlies, and hence the circulation for the year is more easterly than normal. Although this effect problably does contribute to the year to year variations, it seems

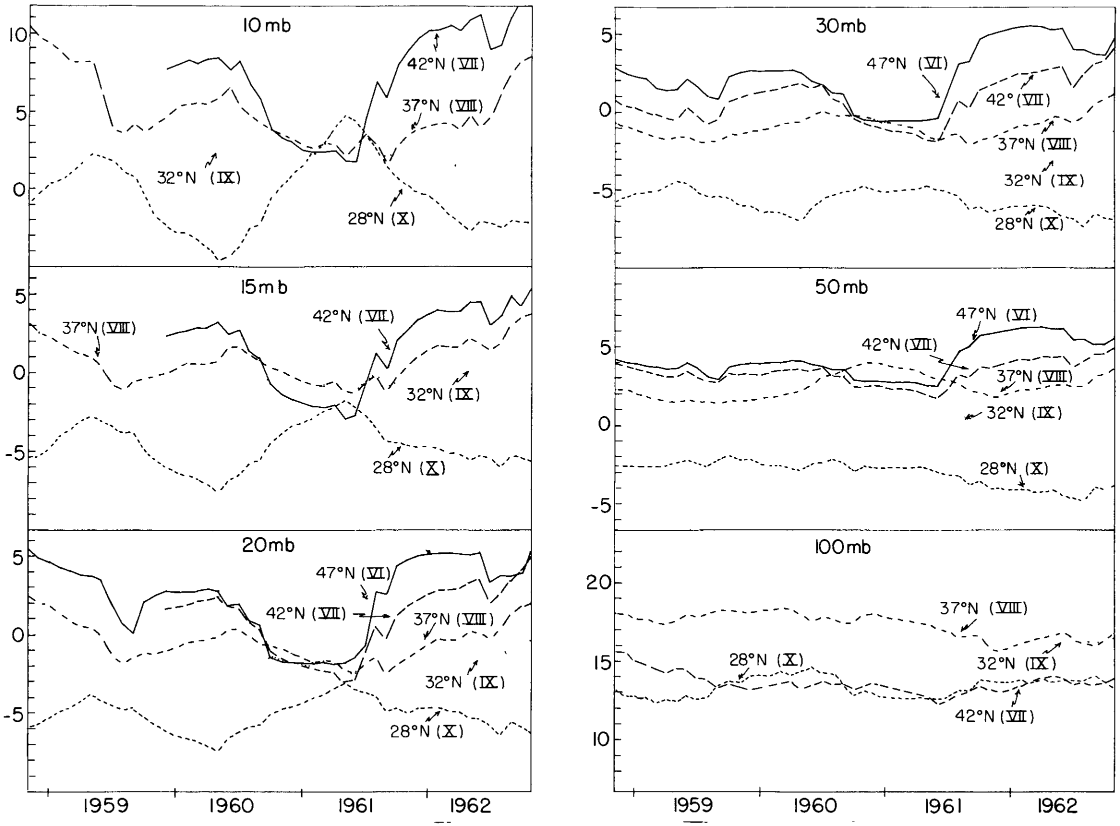


Figure 5.5 Twelve month running means of zonal wind. Units are in m sec^{-1}. The running means are centered on the dates as indicated. December values are placed on the tie marks.

unlikely that it is the major factor in causing them. In the first place, the running means in Fig. 5.6 are not perfect square waves; the transitions between low and high values take place over the course of several months - specifically the winter months, rather than the one month over which the time of the warmings varies. It is also clear from Figs. 5.3 and 5.4 that 1957 was marked by both an early warming and stronger than normal westerlies at mid-latitudes of the southern hemisphere, a combination contradictory to Barbé and Reininger's hypothesis.

Figure 4.1 gives further evidence on how the two hemispheres differ in their response to the biennial oscillation. The section for Christchurch (43°S) shows a marked similarity to that for stations at 20°S (Fig. 3.1) in that the entire 100-20 mb layer is phase locked in the vertical in such a way that the peaks of the oscillation coincide with the winter seasons at all levels. It appears that perhaps this is the only mode in which middle latitudes can respond to the oscillation. It happens that throughout most of the period under investigation, the annual and biennial cycles were phased with respect to one another in such a way that the wind maxima in the tropics coincided with the southern hemisphere winters throughout most of the 10-100 mb layer. Figure 5.6 shows this effect quite clearly.

The situation is effectively reversed in the northern hemisphere. Figure 5.6 indicates that the 20 and 100 mb levels are out of phase with

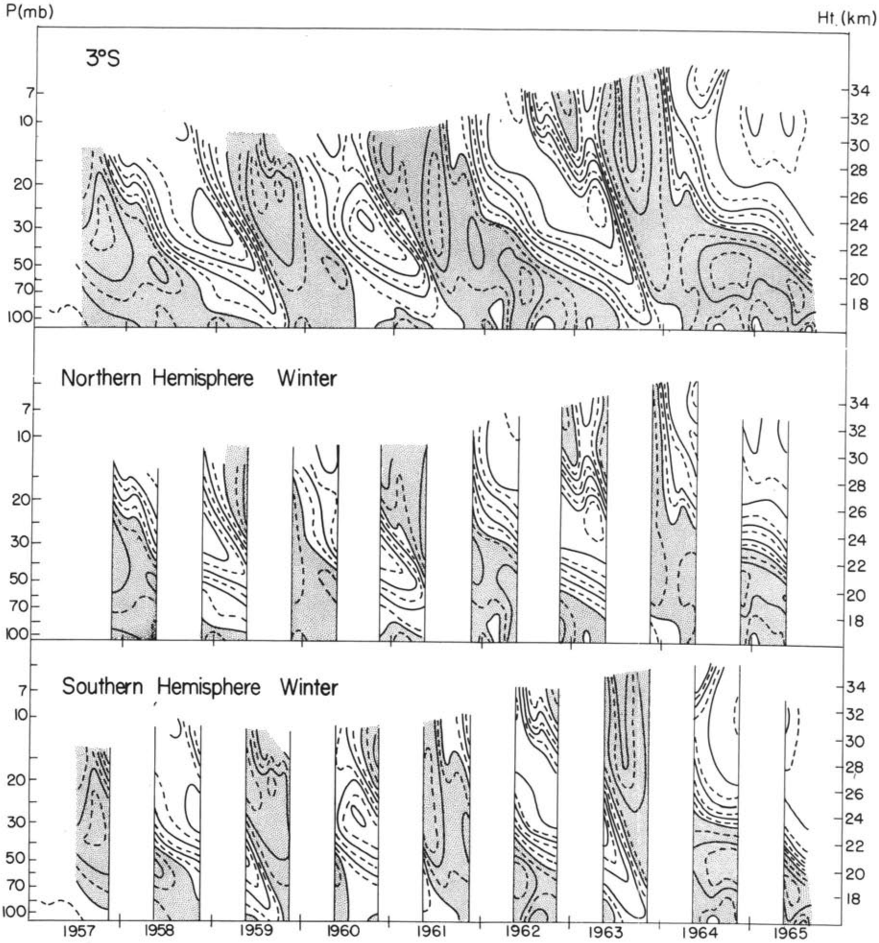


Figure 5.6 Time height sections of zonal wind averaged around 3°S. Lower sections are formed by masking out sections of the upper. The purpose is to view the tropical wind structure during the winter seasons of the two hemispheres. Solid lines are at increments of 10 m sec^{-1}. Westerlies are shaded.

respect to the biennial oscillation during the winter seasons. The middle latitude circulation is apparently unable to respond in this manner, since by 32°N, (Fig. 3.1) the oscillation has virtually disappeared. The winter to winter variation in zonal wind above 30 mb at higher latitudes (Fig. 5.5) may exist for completely different reasons from its southern hemisphere counterpart. Perhaps the vigorous wave disturbances which are characteristic of the northern hemisphere winters above 30 mb are strong enough to influence the strength of the westerlies, just as they apparently influence ozone and temperature. (See Sec. 4.2d). The fact that the winters of 1959-60 and 1961-62, which had particularly strong eddy circulations and stronger than normal poleward momentum transports, were marked by relative easterlies in the subtropics and westerlies at mid-latitudes substantiates the hypothesis. In contrast, the middle latitude wind variations in the southern hemisphere may be brought about mean meridional circulations.

(d) Relationship to the tropical wind variations

Throughout the above discussion it has been tacitly assumed that the year to year differences evident in the middle and high latitude data are associated with the tropical wind fluctuations. Before concluding the discussion it would be well to test the validity of this hypothesis. Perhaps the simplest approach is to test for a correlation between tropical and high latitude events. To do this, one must select some parameter characteristic of the zonal wind field in the tropics to compare with high latitude phenomena. Obviously there is a large number of parameters available to choose from; e. g., the wind at any level, the tendency of the wind at any level, etc. Fortunately, all these parameters are highly correlated with one another at various lag times, so that in reality the choice is not as crucial as it might appear to be.

In her study of northern hemisphere final warmings, Labitzke (1966) selected the tendency of the wind at 20 mb over Canton Is. as the parameter representative of the zonal wind in the tropics. For all the northern hemisphere winters from 1958-65 there proved to be a correspondence between a tendency toward easterlies at Canton Is. and late final warmings. Unfortunately, this particular choice of parameter is not suitable for comparison with southern hemisphere winters since it happens that during those time periods the tendency is usually small or ambiguous.

A more convenient and probably no less relevant parameter proves to be the sign of the vertical shear of the zonal wind between 30 and 10 mb evaluated during the later months of the winter in question at 5° latitude of the winter hemisphere. This can be deduced from Fig. 3. 1 or 3. 4. This quantity is usually not ambiguous and is well correlated with the type of winter warming in both hemispheres. Positive shear - i. e., westerlies above easterlies in the 30-10 mb layer - tends to be associated with early warmings, warm temperatures at the poles, and relative easterlies at middle latitudes, but the correspondence is not perfect. The data in support of this generalization are summarized in Table 5.1.

Thus, the data suggest some interrelation between the parameters under consideration - enough to make it appear unlikely that the correspondence between them is a chance occurrence. However the relation does not seem to be a rigid one, and this makes it difficult and perhaps unwise to speculate on the reasons for it.

5.2 The Biennial Oscillation in the Troposphere

(a) Station data

Symptoms of the biennial oscillation are not so readily apparent in station data in the troposphere. This is understandable in view of the fact that the tropospheric circulation contains a much wider array of eddy sizes than that of the stratosphere and parameters at a given station are largely governed by the positions of the shorter waves which are probably not involved in the biennial cycle. It is indeed rather surprising that in spite of this effectively high noise level, there is evidence of a periodicity in so many tropospheric quantities.

Angell and Korshover (1962, 63) examined wind and temperature data at several tropospheric levels and found some evidence of a periodicity at around 26 months in the data from middle latitudes. However the results are based on a harmonic analysis of data spanning only two cycles of the oscillation, and must therefore be viewed with caution.

Landsberg (1963) and Landsberg et al (1964) have employed power spectrum analysis to study long records of surface data from a large number of stations and have found evidence of an oscillation with the following properties:

1. a period of around 26 months*
2. a tendency for tropical and middle-latitude stations to be out of pha
3. a tendency for phase locking with the annual cycle
4. better organization of the oscillation at some times than at others; during such periods neighboring stations display a more coherent distribution of the phase of the oscillation.

The period of record of the data in Landsberg's study ends in 1950, several years before reliable data became available on the stratospheric oscillation. Hence there is not direct means of verifying that the two phenomena are related. It should be possible to correlate surface data and stratospheric events during more recent years and it is hoped that

*A "period of 26 months" is not meant to imply a strictly periodic compone in the record but rather that in a power spectrum analysis of the data the power in the frequency band centered around 26 months is significantly greater in neighboring frequency bands.

such a study will be undertaken in the near future.*

(b) Circulation indices

The inherent difficulties in tropospheric data, i.e., the fact that a record from one station is not representative of global scale features, can be largely eliminated by combining data from many stations to compute various indices of the global or hemispheric circulation. Shapiro (1964) performed a power spectrum analysis on a time record of the variance of surface pressure and found evidence of a 26 month periodicity. It is significant that the winter months contribute most of the variance to this oscillation. Wagner (1965) arrived at the same result in his study of zonally averaged surface pressure data.

The above studies suggest that there are long period variations in tropospheric data which bear a strong resemblance to those associated with the biennial oscillation in the stratosphere. The data of Miller, Woolf and Teweles (1967) are perhaps the best available evidence that the two phenomena are actually related. The former studied long term

*The most recent decade should prove to be a particularly good interval to examine tropospheric data since during part of this time the stratosphere was marked by a strong two year periodicity. This should be easily detectable in tropospheric data if it is present. Reitschel (1929) describes a world wide two-year oscillation in surface temperatures which prevailed during the interval 1900-12.

variations in the poleward transport of westerly momentum as a function of wave number by performing a Fourier analysis in longitude on daily 500 mb hemispheric maps for a ten year period (approximately 1956-65). Twelve month running means of momentum transports by the first six wave numbers show the following characteristics:

1. There is, throughout most of the decade, a large two year periodicity in the momentum transports.
2. The winter months are almost entirely responsible for this periodicity
3. Wave numbers 1, 2 and 3 show the oscillation most clearly; wave number 2 is out of phase with 1 and 3; i.e., in winters with large northward momentum transports by waves 1 and 3, wave number 2 transports less than a normal amount of momentum northward.

The marked similarities between these year to year variations in 500 mb momentum transports and those previously noted in stratospheri parameters, both within the same period of record, is the strongest evidence thus far presented in favor of a relation between stratosphere and troposphere in the biennial oscillation.

(c) Periodicities in long data records

If it is accepted that the biennial oscillation is perceptible in some large scale meteorological elements in the troposphere such as the positions of the long waves during the winter seasons, then it is not surprising that there are signs of the oscillation in long period records of a variety of more localized parameters. Tables 5. 2 and 5. 3, after Landsberg (1962), give some idea of the widespread occurrence of the phenomenon. It is evident from these tables that the oscillation covers a rather broad scale of frequencies between 2 and 2 1/2 years. The wideness of the frequency band probably results from several factors:

1. The oscillation in the stratosphere is not truly periodic. It apparently undergoes intervals of phase locking with the annual cycle, when its actual period is exactly two years, and other intervals, such as 1963-65, when its period is variable and in excess of two years. (Sec. 4. 3a).

2. Apart from the annual cycle, the noise (spoken of in the context of the discussion of Sec. 5. 2a) inherent in tropospheric data accounts for most of the energy present in the power spectrum. For an infinitely long record of data this energy would fall into a spectrum whose frequency distribution would be featureless in the neighborhood of two years. However, for the finite records of data under consideration, the spectrum of this noise has features which are capable of shifting or even obliterating the spectral peak associated with the biennial cycle.

Table 5. 2 (After Landsberg (1962)).

The 2 to $2^1/_2$ year rhythm as indicated in various records of atmospheric elements

Locality	Element	Interval	Rhythm	Author	Year publ.
W., C., NE. U. S.	Temperature	var. 1839—1873	2.1	CLAYTON [5]	1884
Russia	Ice, snowcover	1874—1893	2.0	WOEIKOF [9]	1895
Stockholm	Winter temp.	1757—1906	2.0	WOEIKOF [10]	1906
Stockholm	Temperature	1759—1912	2.0—2.3	WALLÉN [12]	1913/14
Uppsala	Precip.	1836—1912	1.9—2.2	WALLÉN [12]	1913/14
Bergen	Temp., press., precip.	1861—1914	2.0—2.3	BIRKELAND [14]	1916
Bathurst, W. Afr.	Precip.	not given	2.1	BROOKS [25]	1920
10 German Sta.	Variab. temp.	1870—1919	2.0	BAUR [21]	1922
German Sta.	Temperature	1870—1919	2.0	BAUR [21]	1922
10 Bavar. Sta.	Precip.	1880—1922	2.4	BAUR [22]	1923
2 Bavar. Sta.	Pressure	1884—1919	2.4	BAUR [22]	1923
Azores, Iceland	Pressure	1884—1920	2.25—2.4	BAUR [23]	1924
Capetown, Bombay	Pressure	1884—1922	2.25	BAUR [23]	1924
Europe, U. S.	Temperature	18th cent. to 1922	2.0—2.5	CLOUGH [19]	1924
Gt. Britain	Var. elements	18th cent. to 1922	2.2—2.4	BAXENDELL [28]	1925
Punjab	Precip.	1863—1918	2.1	ALTER [17]	1926
U. S. Pac. Coast	Precip.	1850—1922	2.4	ALTER [18]	1927
Milan, Padua	Precip.	1764—1863	2.2	BRUNT [29]	1927
Edinburgh	Precip.	18th & 19th cent.	2.1, 2 4	BRUNT [29]	1927
Edinburgh	Pressure	,, ,,	2.4	BRUNT [29]	1927
Europ. Cont.	Temperature	,, ,,	2.1—2.2	BRUNT [29]	1927
Paris	Pressure	,, ,,	2.2—2.4	BRUNT [29]	1927
Berlin	Temperature	1822—1921	2.2	BAUR [24]	1927
Zwanenburg	Temperature	1743—1760	2.3	BAUR [24]	1927
Styckisholmur	Pressure	1890—1907	2.3	BAUR [24]	1927
Styckiskolmur	Pressure	1908—1925	2.2	BAUR [24]	1927
Up. Rhenanian	Precip.	1906—1927	2.2	BAUR [24]	1927
Whole earth	Dec.-Feb. temp.	1901—1912	2	RIETSCHEL [34]	1929
North Atl. Sta.	Precip.	1849—1929	2.3	VISSER [36]	1937
State College, Pa.	Temp. range	1886—1938	2.4	CONRAD [35]	1940
Philadelphia, Pa.	Temperature	1825—1930	2.0	LANDSBERG [7]	1941
Batavia	Pressure	1866—1932	2.3	BOER [32]	1941
So. Oscillation	Press. grad.	1866—1940	2.3	BERLAGE [38]	1956
Woodstock, Md.	Temperature	1870—1956	2.1	LANDSBERG et al. [8]	1959
So. Oscillation	Press. grad.	1866—1940	2—2.5	BERLAGE and de BOER [40]	1960
SW U. S.	Precip.		2—3	SELLERS [48]	1960
Djakarta	Pressure	1866—1960	2—2.5	BERLAGE [41]	1961
Eq. Stratosphere	Wind dir.	1954—1959	1.9—2.4	VERYARD and EBDON [43]	1961
Eq. Stratosphere	Wind dir.	1954—1960	2.2	REED and ROGERS [44]	1962

Table 5.3 (After Landsberg (1962)).

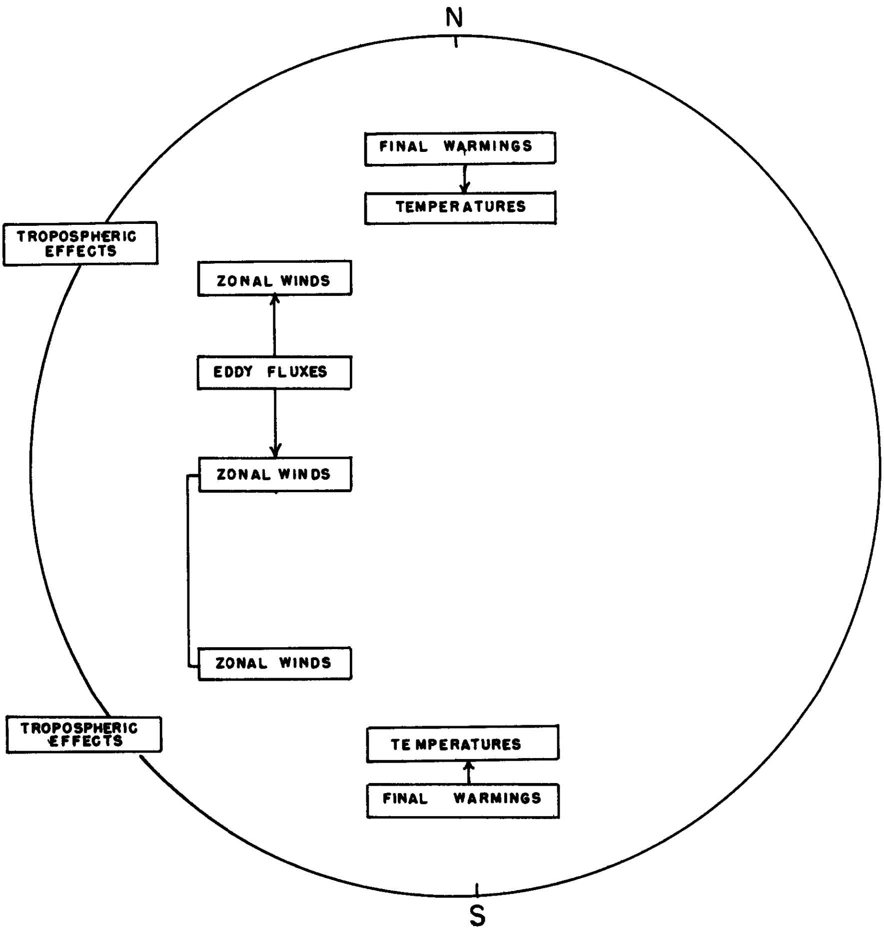
Indications of 2 to $2^1/_2$ year rhythm in other than meteorological time series

Locality	Element	Interval	Rhythm	Author	Year publ.
Norweg. Coast	Winter sea surf temp.	1874—1904	2.0	PETTERSSON [11]	1905
Sweden	Lake levels	1774—1912	2.5—2.7	WALLÉN [12]	1913/14
North Atlantic	Sea surf. temp.		(2)	HELLAND-HANSEN and NANSEN [15]	1917
Earth	Sunspots	1750—1922	2.3	CLOUGH [19]	1924
Earth	Solar constant	1918—1926	2.1	ABBOT [49]	1927
Nile	Floods	641—1451	2.0	BROOKS [26]	1928
N & S. Am. India Scandinavia	Varves	18,000 yrs.	2.2	BROOKS [27]	1928
Earth	Sunspots	1750—1934	2.2	DOUGLAS [50]	1936
Var. geol. format.	Varves		2—2.7	ANDERSON [46]	1961
West. U.S.	Tree rings	3,000 yrs.	2.1, 2.7	BRYSON and DUTTON [47]	1961

3. Because of interactions between the atmosphere and oceans on a time scale of years, a certain amount of persistence from one year to the next is to be expected in meteorological data. The effect of this persistence on the frequency distribution of disturbances in the atmosphere is to slightly reduce the amplitude of short period disturbances relative to longer period ones. This differential amplification of disturbances should be most noticeable for periods on the order of the time scale of the atmosphere ocean interactions. If the southern oscillation is typical of the time scale of such interactions, then this effect should be felt most strongly for periods in the order of several years. Nordo (1965) has noted that if there were a relative maximum at two years in the frequency spectrum of a meteorological variable, the effect of persistence would be to lengthen the apparent period of the peak. The amount of this "red shift" would be determined by the strength of persistence relative to the sharpness of the peak. This could vary considerably from one meteorological parameter to another and from one geographical location to another. Hence this effect might explain why meteorological variables exhibit a range of periods between 2 and 2 1/2 years rather than a single period at 2 years.

5.3 Summary

Figure 5.7 depicts in schematic form the principal results of this chapter. Each of the boxes represents some phenomenon believed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FINAL WARMINGS
TEMPERATURES
TROPOSPHERIC EFFECTS
ZONAL WINDS
EDDY FLUXES
ZONAL WINDS
ZONAL WINDS
TROPOSPHERIC EFFECTS
TEMPERATURES
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Figure 5.7

to be associated with the biennial oscillation. Wherever phenomena in different regions are related to one another by a known physical mechanism, the boxes are connected. The connecting lines are replaced by arrows in cases where the relationship is suggestive of cause and effect.

At present our understanding of the modes in which circulations at different levels and latitudes interact with one another is far from adequate for describing this complex network of interactions involving virtually the entire atmosphere. Thus it is not surprising that this chapter has fallen short of providing a comprehensive explanation of all the observed features of the biennial oscillation. Each missing link in the figure is worthy of an involved study in itself. To speculate on possible modes of interaction without undertaking such a study would not do justice to the subject and would tend to confuse those results for which there is a strong physical basis with those which are mere speculation. Perhaps the blocks should best be left conspicuously unconnected as suggestions for future studies.

CHAPTER VI

LONG PERIOD VARIATIONS AS A GLOBAL PHENOMENON

Chapters III, IV, and V have dealt with various aspects of the long period fluctuations in the atmosphere associated with the "biennial" or "26 month" oscillation. The apportioning of the subject material among the three chapters was determined largely by convenience in discussing cause and effect relationships. The purpose of this final chapter is to view the fluctuations as a global phenomenon, thus integrating the results obtained in the earlier chapters. We will begin with a brief summary of these results.

6.1 A Review

Chapter III discussed the most noticeable and well documented features of "the biennial oscillation", i.e., the long period wind and temperature fluctuations in the tropical stratosphere. It was shown that long period variations in the eddy momentum fluxes are instrumental in producing the sequence of alternating easterly and westerly wind regimes in the tropics. It was argued that a mean downward drift is responsible for the propagation of these regimes into the lower stratosphere, and that small variations in this mean downward motion give rise to the temperature fluctuations which are necessary to keep the wind and temperature fields in geostrophic balance. Thus, the inference was that both the observed wind and temperature variations in the tropical stratosphere are a response to variations in the eddy fluxes of momentum into and out of the region.

Chapter IV was an attempt to determine the nature and cause of the variations in momentum transport, with hopes of coming one step closer to the cause of the "oscillation". It was shown that at least in the northern hemisphere the variations in momentum transports are but one symptom of an overall modulation in the intensity of the eddy activity at middle latitudes. Since eddy activity is virtually confined to the winter season, this represents specifically a winter to winter variation. The winter temperatures and total ozone content at middle latitudes of the northern hemisphere were shown to exhibit year to year changes consistent with this modulation of eddy activity for the five winters for which there is data on both quantities. On this basis it was inferred, from a longer record of ozone data, that there is a tendency for a two year periodicity in the intensity of the winter circulations. It was suggested that such a periodicity could arise within the atmosphere from an interaction between the eddy circulations at middle latitudes and the mean circulations in the tropics. Since it was shown in Chapter III that the eddy circulations drive the mean circulations of the tropics, it was only necessary to demonstrate that the mean circulations, in turn, exert a feedback upon the eddy circulations, to prove that such an interaction does exist. Three possible schemes for providing the necessary feedback were discussed: (1) a modulation by the mean zonal flow of the eddy energy coming into the region from below, (2) the influence of the horizontal shear in the mean zonal flow on the shape of the eddies and (3) the effect of the mean zonal flow at low latitudes upon the stability characteristics of the polar vortex.

Chapter V was devoted to some of the many manifestations of the long period variations which were not discussed in the previous chapters. Within the stratosphere, year to year variations in polar temperatures, final warming dates, and zonal winds at middle latitudes were discussed. It was suggested that polar temperatures are merely reflections of the timing of the final warmings. There appeared to be a basic difference in the character of the wind variations at middle latitudes in the two hemispheres, with those in the southern hemisphere being closely related to tropical wind regimes, and those in the northern hemisphere being more closely identified with the temperature and ozone variations at those latitudes which are believed to result from a modulation of eddy activity. Periodicities in tropospheric quantities were discussed and evidence was presented for a relation between troposphere and stratosphere in these long period variations.

These results are summarized in Fig. 6.1 which is a composite of Figs. 3.6, 4.9 and 5.7. The reasons for the connections between tropospheric and stratospheric effects will be discussed in Sec. 6.4.

6.2 Interhemispheric Differences

It will not be possible to determine definitely whether any basic differences between hemispheres exist in these long period variations until data on eddy fluxes become available for the southern hemisphere. The data presently available show evidence of some distinct differences in other, less basic quantities such as zonal winds at middle latitudes (Sec. 5.1c) and temperatures at middle and high latitudes (Sec. 5.1a). It is difficult to explain even these dissimilarities without allowing for some basic differences in the dynamical processes taking place in the two hemispheres.

It is certainly not surprising that differences should exist in the winter circulations of the two hemispheres. It was shown in Sec. 4.1a that the northern hemisphere winter exhibits a much stronger short term variability than its southern hemisphere counterpart. This is indicative of more vigorous disturbances within the northern hemisphere vortex. In view of this fact, it might be expected that the northern hemisphere circulation would exert the dominant influence on the mean flow in the

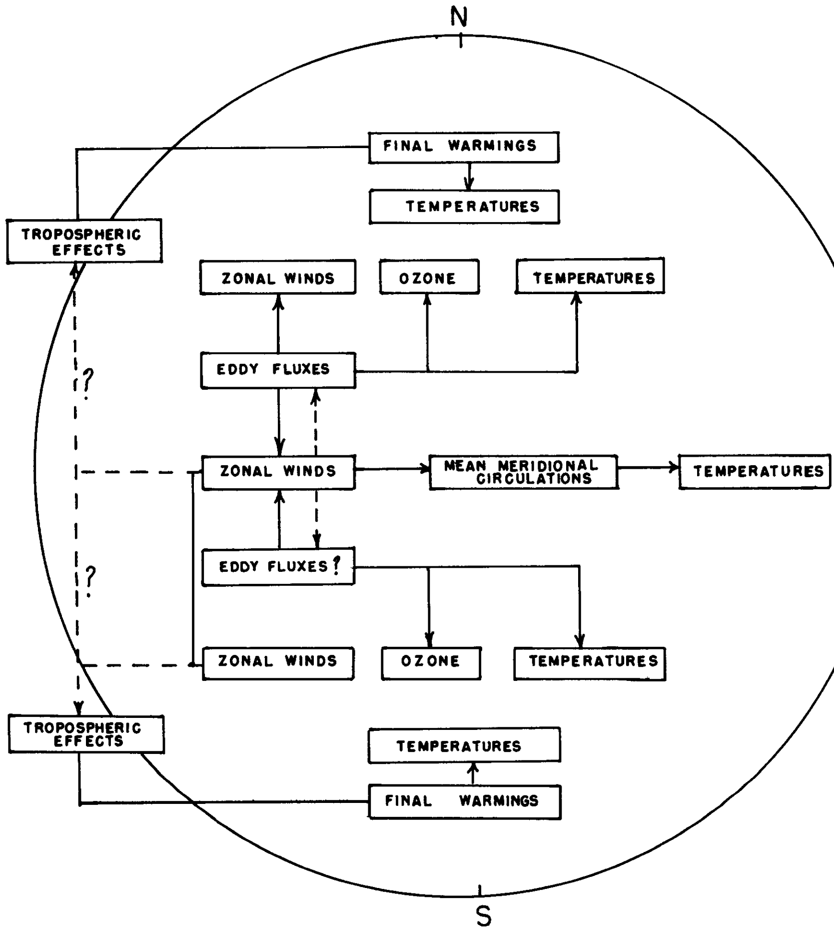
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Figure 6.1

tropics. The time-height section for $3^{o}S$ (Fig. 3.1b) suggests that above 30 mb, where the zonal accelerations are chiefly a response to the eddy transports, the large accelerations usually take place during the northern hemisphere winters, which lends support to this hypothesis.

If there is any essential difference in the roles which the two hemispheres play in these long period variations (i.e., if the observed differences are not accidental features of the short record of data presently available) then, as Tucker (1966) has pointed out, it is likely that the troposphere is deeply involved in the dynamics of the phenomenon. For presumably, basic interhemispheric differences in the general circulation arise from the different land-sea distributions in the two hemispheres and this effect is felt primarily in the troposphere.

6.3 The Role of the Troposphere

Newell (1964b) suggested that, inasmuch as the tropospheric heat engine at middle latitudes is the source of the energy for the motions of the lower stratosphere, it would be well to look to that region for possible causes of the biennial oscillation. From an energy standpoint this region is certainly more appealing than the upper atmosphere since its mass is two orders of magnitude larger and consequently "the tail doesn't have to wag the dog", so to speak.

The discussion of long period variations in tropospheric parameters in Sec. 5.2 suggests that the troposphere is indeed involved in the "biennial oscillation", at least to a limited extent. However, as appealing as the idea may be, there are a number of serious difficulties involved in viewing the troposphere as the self contained source of the long period variations. The five year study of Krueger, Winston and Haines (1965) fails to show evidence of any significant year to year variations in the transformations between potential and kinetic energy in the zonal and eddy forms, (There are significant winter to winter variations in the same quantities in the stratosphere). A further difficulty is that the troposphere, of itself, lacks the capacity for remembering from one winter to the next, as in the stratosphere, where the mean zonal flow in the tropics performs the functions of a memory (Sec. 4.3c). Teweles (1965) suggested that an interaction with ocean currents might serve as a memory for the troposphere, but surface temperatures and pressures for a number of ocean stations during the past decade fail to show any evidence of long period variations of the type occurring at higher levels.

As an alternative to a memory, as such, Newell (1964a) suggested that long period variations in incoming visible solar radiation might produce fluctuations in the intensity of the tropospheric heat engine and that these could, in turn, be transmitted to the stratosphere. This explanation also meets with difficulty, for the data of Miller et. al. (1967)

and Krueger, Winston and Haines (1965) cited above, argue in favor of variations in rather specific aspects of the tropospheric circulation, as opposed to a modulation of the overall intensity of the tropospheric heat engine.

The fact that the long period variations in question are primarily due to winter to winter differences is, in itself, a clue as to their origin. Although the tropospheric circulation is undoubtedly more vigorous in winter than in summer, the difference between seasons is only a matter of degree. This is in contrast to the stratospheric circulation (see Sec. 4.1) in which virtually all the eddy activity is confined to the winter season. Now the fact that the year to year differences, even in tropospheric quantities such as the 500 mb momentum transports, can be attributed almost entirely to the winter seasons suggests that the stratosphere, rather than the troposphere, is the source region.

6.4 A Further Examination of Possible Causes

In Sec. 4.3c we discussed a specific class of mechanisms which might be responsible for producing the long period variations. At that time it was not possible to consider the question of whether such mechanisms could account for all the observed manifestations of the "biennial oscillation", since high latitude and tropospheric effects had not yet been discussed. At this point, having considered the long period variations in these regions, we return to this question.

The mean zonal flow in the tropics determines the outer boundary condition on the polar vortex, (Sec. 4.3c) and the upper boundary condition on a considerable portion of the troposphere. It was suggested in Sec. 4.3c) that changes in the lateral boundary condition of the polar vortex might have an important effect on the stability of the vortex, and thus it might influence the timing of warmings. Similarly, if wave disturbances in the troposphere are sensitive to the zonal wind structure at this upper boundary then it may be possible for the long period variations within the stratosphere to induce variations in the structure of the waves in the troposphere.

As an example of such a boundary effect, we might consider the "work term", given by the correlation between pressure and vertical velocity on a horizontal surface at this boundary. This term represents the main contribution to the energy flow through this boundary (Miller, 1966). If, as Charney and Drazin's (1961) results suggest, and the interlevel correlations (Sec. 4.3c) support, the zonal flow does affect the vertical propagation of wave energy, then it may impose limitations on the size of pressure or vertical motion variations on this surface, or the correlation of the two quantities in wave disturbances.

There is also the possibility of direct coupling between the waves in stratosphere and troposphere in the large winter disturbances. Julian and Labitzke (1965) and Labitzke (1966) have shown evidence that stratospheric warmings are related to blocking in the troposphere.

In postulating that the stratosphere contains the mechanism for producing the long period variations there is no intention of implying that energy flows from stratosphere to troposphere. On the contrary, the mechanisms suggested are consistent with and one of them actually depends upon the upward propagation of energy from troposphere to stratosphere. The implication is that the stratosphere is capable of regulating the amount of this energy coming into the region from below.

6.5 Concluding Remarks

The purpose of this thesis has been to examine the long period variations in the atmosphere from an observational standpoint, using the simplest possible analysis techniques and making no assumptions regarding periodicities, with hopes of gaining some insight into the cause of the phenomenon.

It was found that the zonal wind data presently available in the tropics are adequate for defining these variations in considerable detail. The variations, though not strictly periodic at all times, do exhibit a very simple behavior which is remarkable in its consistency with respect to both latitude and longitude. Consideration of the momentum, heat and energy budgets relevent to these motions led to the conclusion that they are driven by quasi-horizontal exchange processes involving the extratropical circulation of the winter hemisphere. This result is in agreement with the contention of Charney (1963) that "in the absence of condensation, tropical motions ...... [are] driven primarily by lateral coupling with extratropical and precipitating tropical motions".

In light of these findings it appears extremely unlikely that temporal variations in radiative heating could be the cause of the oscillation. Future work towards an explanation of its occurrence would do well to consider the question of year to year variations in the winter stratospheric circulations.

It is hoped that the data presented herein will be useful to those interested in modeling the stratosphere and that the discussion will be influential in shaping future models of "the 26 month oscillation".

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APPENDIX

The discussion of data sources and data processing techniques contained in the main text is probably adequate for the reader interested in this work for the sake of the scientific results contained therein. The following, more detailed discussion is intended more as a help to those contemplating observational studies along the same lines as this one. The purpose is to acquaint the reader with the basic sources of radiosonde data for the stratosphere, particularly in the tropics, and to discuss some of the problems involved in making computations of the type done in this thesis.

A-1 Data Sources

The daily data used in this thesis were taken from three general sources:

(a) The tapes of the MIT General Circulation Data Library, containing data for 249 northern hemisphere stations, plus Canton Island for the five year period May 1958-April 1963. The reporting levels are 100, 70, 50, 30, 20, 15 and 10 mb, and the hour is 00Z only.

These data were originally extracted from Card Decks 524-525 and 545-645, in storage at the National Weather Records Center (NWRC), Asheville, N.C. These decks constitute the information contained in the Northern Hemisphere Daily Bulletins, published by NWRC. The former decks contain data for foreign stations, the latter for U.S. operated stations, which account for more than 90% of the stratospheric data available in the northern hemisphere during this period.

The locations of the stations are shown in Fig. A, and their names and other pertinent information are given in Table A-1. The U.S. operated stations have WABAN numbers in the table. Table A-2 gives the number of individual daily observations during the five year period which contain both wind and temperature, as a function of latitude belt and level. (70 mb is not included because it did not become a regular reporting level until the beginning of 1961).

A small amount of supplementary data for 12Z and levels above 10 mb was extracted directly from the Northern Hemisphere Bulletins. This was used to extend Figs. 3.1 and 3.4 to higher levels.

(b) In order to extend the period of record for tropical stations, data for the periods July 1957-April 1958 and May 1963-December 1964 for all the U.S. operated stations south of 25°N in Table A-1 were purchased from NWRC, in the form of Card Decks 545 and 645. Both 00Z and 12Z data are included, and reporting levels include 80, 60, 40, 25, 7 and 5 mb in addition to those listed in Item (a).

Items (a) and (b), taken together constitute an unbroken 7½ year data sample for about 40 northern hemisphere tropical stations. These data are being collected on a continuing basis at MIT, and an additional half year is available at the time of this writing.

(c) Southern hemisphere daily data were from the data centers of several countries:

1. Card Decks 545-645 were purchased from NWRC. These contain data for U.S. operated stations in South America north of 25°S and Ascension Is. The period of record is July 1957-December 1963.

2. Data for all Australian stations north of 25°S were purchased on cards from the Bureau of Meteorology of Australia. Reporting levels used were 60, 70, 80 and 90 thousand feet; the period of record is July, 1957-December, 1964.

3. Data for South African stations north of 25°S were purchased on cards from the Weather Bureau of the Union of South Africa. The period of record is July 1957-December 1960. Reporting levels are 70, 50 and 30 mb.

4. Data for Portugese colonies in Africa were provided without charge by the Servico Meteorologico Nacional of Portugal in the form of tables. These were punched onto cards at MIT. The period of record is January, 1961-December, 1962. Reporting levels are 100, 70, 50, 30 and 20 mb. In general, these data consist of once daily observations at various hours.

At the present time it does not appear that there is enough southern hemisphere data in the stratosphere to compute reliable momentum and heat flux statistics and therefore the only use made of it is in the monthly mean zonal winds. Table A-3 contains pertinent information on the stations in this section.

In addition to daily data, a small amount of monthly mean zonal wind data was used in the sections. Mr. J. Korshover of ESSA provided the 1965 data for Canton Island, (Fig. 3.1) and the New Zealand Meteorological Service provided the data for Christchurch, N.Z., used in Fig. 4.1. The mean monthly maps published in Meteorologische Abhandlungen were used as a basis for the standing eddy computations.

A-2 Preprocessing

Momentum flux statistics are extremely sensitive to any type of error in the daily data which produces abnormally large wind components. For instance, in the tropics, where the meridional wind component is usually very small, an erroneous wind direction can result in a conspicuously large computed value of momentum transport for the entire month at a given station.

In order to prevent contamination of the flux statistics by errors of this type, the wind data were edited by subjecting the values of the wind components to certain predetermined limits. These limits were arrived at by computing the mean zonal component, and the temporal standard deviations of both components as a function of latitude, level, and season. The zonal component was allowed to vary within a range of about 4 standard deviations from the appropriate mean, and the meridional component to vary within about 4 standard deviations from zero. The rejected winds were inspected and care was taken to see that an excessive number of rejections did not occur because of inappropriate limits. Rejection rates ranged from about a tenth of a percent most months to almost a percent during some of the very active winter months. A rough hydrostatic check was also applied to all the daily temperature data used in this study.

This method of editing data does not, by any means, remove all errors, nor is it possible to keep it from rejecting some correct data. However, with appropriate limits it does eliminate those errors most injurious to the eddy flux statistics without screening out an unduly large amount of good data. These preprocessing procedures undoubtedly bias the flux statistics toward lower values. However, in view of the small rejection rates, it is doubtful that this effect is even noticeable, compared with the natural bias due to the difficulty in tracking balloons in high wind speeds. And since the statistics derived from these data are used more in a qualitative than in a quantitative manner, this type of bias does not cause serious objections.

A-3 Calculation of the Transient Eddy Statistics

Monthly mean values, $\overline{(\;)}$, were computed at each station for the following quantities.

u	zonal wind component
v	meridional wind component
$\overline{u'v'}$	northward momentum transport by transient eddies
$\overline{v'T'}$	northward heat transport by transient eddies
σu	temporal standard deviation of the zonal wind component
σv	temporal standard deviation of the meridional wind component

Station values of each of the above quantities except $\overline{v}$ were then combined so as to obtain average values $[(\;)]$, for each latitude belt. The station values were weighted by the number of reports for the month so that, for instance, a station which reported on twenty days of the month would be ten times as influential as a station which reported only twice. This procedure does not recognize the fact that a small number of observations spaced far apart in time within a month are more representative than the same number of observations if they should happen to occur on consecutive days. However this is probably not a serious objection, since such ambiguities arise only in situations where the data are sparse and these are given relatively little weight. A more serious objection is that strictly speaking, a zonal average should be derived by drawing maps for each quantity at each level and combining values at equally spaced grid points. Obviously, that is impossible in this case, due to the absence of data over large portions of each latitude belt. A slightly cruder method which also pays heed to longitudinal representation would be to weight each station by the longitudinal extent which its data represent, so that stations in regions of sparse data would be more influential than those which are densely spaced. However if isolated stations are used to represent the vast areas which surround them large errors will still result, since over the space of a few hundred miles, conditions can change markedly. While the method used in this study fails to produce mean quantities which are uniformly representative with respect to longitude, it does have the distinct advantage than random errors are minimized by the equal weighting of all observations.

The above arguments do not fully justify what has been done. The results are valid if and only if the time variations in the quantities derived by this method are representative of the time variations in the true zonal means. There is reason to hope that this may be the case for the parameters used in this study. Buch (1954) noted that latitudinal means of heat and momentum transports as calculated by averaging together station data were almost identical to those obtained from map analyses. It is hoped that future studies with more complete data coverage will bear this out. For the present, the latitudinal means derived in this study can be said to apply mainly to the North American sector, from which the great majority of the data are taken.

The mean meridional wind $[\bar{v}]$, and temperature, $[\bar{T}]$, have not been computed by this method. The former quantity is very small compared to typical station values within the latitude belt while the latter is strongly influenced by radiation corrections at individual stations which are in turn a function of solar elevation angle, and consequently of longitude. Hence these terms are extremely sensitive to the longitudinal distribution of stations.

Of the standard deviation terms σ_u and σ_v , the latter is more meaningful as an index of the intensity of the transient eddies. A trend in the zonal wind within a month can contribute significantly to the value of σ_u , and therefore this term does not lend itself readily to physical interpretation. Monthly trend is less of a problem with σ_v , since the range of $\bar{v}$ is much less than that of $\bar{u}$. For this reason, σ_v data is more heavily stressed.

A-4 Calculation of the Standing Eddy Statistics

Standing eddy statistics were derived from the monthly mean geopotential height fields by computing the wind components at gridpoints, using the finite difference forms of the geostrophic equation. From these, the zonal average statistics were computed directly. Centered finite differences were used with grid points at 20^{o} intervals in longitude and 5^{o} in latitude.

Table A-1. List of stations (northern hemisphere).

Sta No.	Trav. No.	WMO No.	WABAN No.	Lat. Bd	Name	Lat.	Long.
1	000	01001		1	Jan Mayen	71N	08W
2	003	01028		1	Bjørnøya	75	19E
3	005	01152		2	Bodo	67	14
4	006	01241		3	Orlandet	64	10
5	007	01384		3	Garder Moen	60	11
6	008	01415		4	Stavanger	59	06
7	012	02836		2	Sodankyla	67	27
8	013	02963		3	Jokioinen	61	23
9	025	03953		5	Valentia	52	10W
10	026	04018	16201	2	Keflavik	64	23
11	027	04202	17605	1	Thule A. B.	77	69
12	040	07110		6	Brest (Guipavas)	48	04
13	041	07145		6	Trappes	49	02E
14	043	07180		6	Nancy	49	06
15	044	07354	34048	6	Chateauroux	47	02
16	045	07510		6	Bordeaux	45	01W
17	046	07645		7	Nimes	44	04E
18	052	08509	13201	8	Lajes	39	27W
19	055	08594		12	Sal	17	23
20	058	10202		5	Emden	53	07E
21	065	10739		6	Stuttgart	49	09
22	066	10866		6	Munchen	48	12
23	067	11035		6	Wien	48	16
24	075	13130		6	Zagreb	46	16
25	076	13276		6	Beograd	45	21
26	077	13334		7	Split	44	16
27	080	15614		7	Sofia Observ.	43	23
28	090	16716		8	Athinai	38	24
29	296	40181		9	Beer Yaaqov	32	35
30	297	40427		10	Bahrain	26	51
31	326	45004		11	Hong Kong	22	114
32	332	47122	43242	8	Osan	37	127
33	334	47187	43263	9	Mosulpo	33	126
34	337	47420		7	Nemuro	43	146
35	339	47582		8	Akita	40	140
36	340	47590		8	Sendai	38	141
37	341	47600		8	Wajima	37	137
38	342	47646		8	Tateno	36	140
39	343	47678		9	Hachijojima	33	140
40	344	47744		8	Yonago	35	133

Sta No.	Trav. No.	WMO No.	WABAN No.	Lat. Bd	Name	Lat.	Long.
41	346	47778		9	Shionomisaki	33N	136E
42	347	47807		9	Fukuoka	34	130
43	348	47827		9	Kagoshima	32	131
44	349	47909		10	Naze	28	130
45	350	47931	42204	10	Kadena Okinw.	26	128
46	351	47963		10	Torishima	30	140
47	355	48694		13	Singapore	01	104
48	358	48900		12	Saigon	11	107
49	483	60020		10	St. Cruztenri	28	16
50	484	60119	13017	9	Kenitra	34	07
51	485	60390		8	Alger	37	03
52	486	60570		9	Colomb-Bechar	32	02W
53	488	60625		10	Aoulef	27	01E
54	490	61052		12	Niamey	13	02
55	491	61401		10	Fort-Trinquet	25	12 W
56	492	61642		12	Dakar (Yoff)	15	17
57	493	62011	33123	9	Tripili	33	13E
58	495	62062		9	Tobruk	32	24
59	496	62306		9	Mersa Matruh	31	27
60	497	62366		10	Cairo	30	31
61	498	62378		9	Helwan	30	31
62	499	62414		11	Asswan	24	33
63	500	62721		12	Khartoum	15	33
64	502	64650		13	Bangui	04	19
65	503	64700		12	Fort-Lamy	12	15
66	505	65201		13	Lagos (Ikeja)	07	03
67	506	65578		13	Abidjan	05	04W
68	507	70026	27502	1	Barrow	71	157
69	508	70086	27401	2	Barter Island	70	144
70	509	70133	26616	2	Kotzebue	67	163
71	510	70200	26617	2	Nome	65	165
72	511	70219	26615	3	Bethel	61	162
73	512	70231	26510	3	McGrath	63	156
74	513	70261	26411	2	Fairbanks	65	148
75	514	70273	26409	3	Anchorage	61	150
76	515	70308	25713	4	St. Paul Island	57	170
77	516	70316	25624	4	Cold Bay	55	163
78	517	70326	25503	4	King Salmon	59	157
79	518	70350	25501	4	Kodiak	58	153
80	519	70361	25339	4	Yakutat	60	140

Sta No.	Trav. No.	WMO No.	WABAN No.	Lat. Bd	Name	Lat.	Long.
81	520	70398	25308	4	Annette Island	55N	132W
82	522	70414	45715	5	Shemya	53	174E
83	523	70454	25704	5	Adak	52	177W
84	524	72201	12850	10	Key West	25	82
85	525	72202	12839	10	Miami	26	80
86	526	72206	13889	10	Jacksonville	30	82
87	527	72208	13880	9	Charleston	33	80
88	528	72211	12842	10	Tampa	28	83
89	529	72221	13858	9	Valpariso	30	87
90	530	72222		10	Pensacola	30	87
91	531	72226	13895	9	Montgomery	32	86
92	532	72232	12863	10	Burrwood	29	89
93	533	72235	13956	9	Jackson	32	90
94	534	72240	13941	10	Lake Charles	30	93
95	535	72248	13957	9	Shreveport	32	94
96	536	72250	12919	10	Brownsville	26	97
97	537	72251	12926	10	Corpus Christi	28	97
98	538	72253	12921	10	San Antonio	30	98
99	539	72259	13911	9	Fort Worth	33	98
100	540	72261	22001	10	Del Rio	29	101
101	541	72265	23023	9	Midland	33	102
102	542	72270	23044	9	El Paso	32	106
103	543	72273		9	Ft. Huachucha	32	110
104	544	72274	23160	9	Tucson	32	111
105	545	72280		9	Yuma	33	115
106	546	72290	03131	9	San Diego	33	117
107	547	72295	93197	9	Los Angeles	34	118
108	548	72304	93729	8	Hatteras	35	77
109	549	72308	13737	8	Norfolk	37	76
110	550	72311	13873	9	Athens	34	83
111	551	72317	13723	8	Greensboro	36	80
112	552	72327	13897	8	Nashville	36	87
113	553	72340	13963	8	Little Rock	35	92
114	554	72353	13967	8	Oklahoma City	35	98
115	555	72363	23047	8	Amarillo	35	102
116	556	72365	23050	8	Albuquerque	35	107
117	557	72374		8	Winslow	35	111
118	558	72385	03133	8	Yucca Flat	37	116
119	559	72386	23169	8	Las Vegas	36	115
120	560	72392		8	Vandenberg AFB	35	121

Sta No.	Trav. No.	WMO No.	WABAN No.	Lat. Bd	Name	Lat.	Long.
121	561	72394	23273	8	Santa Maria	35N	120W
122	562	72403		8	Washington	39	77
123	563	72405	93734	8	Washington, DC	39	77
124	564	72425		8	Huntington	38	83
125	565	72429	13840	7	Dayton	40	84
126	566	72445	13983	7	Columbia	39	92
127	567	72451	13985	8	Dodge City	38	100
128	568	72456	13996	7	Topeka	39	96
129	569	72469	23062	7	Denver	40	105
130	570	72476	23066	7	Grand Junction	39	109
131	571	72486	23154	7	Ely	39	115
132	572	72493	23230	8	Oakland	38	122
133	573	72506	14756	7	Nantucket	41	70
134	574	72518	14735	7	Albany	43	74
135	575	72520	94823	7	Pittsburgh	41	80
136	576	72528	14733	7	Buffalo	43	79
137	577	72532	14842	7	Peoria	41	90
138	578	72553	94918	7	Omaha	41	96
139	579	72562	24023	7	North Platte	41	101
140	580	72572	24127	7	Salt Lake City	41	112
141	581	72576	24021	7	Lander	43	109
142	582	72583	24128	7	Winnemucca	41	118
143	583	72597	24225	7	Medford	42	123
144	584	72600		7	Sable Island	44	60
145	585	72606	14764	7	Portland	44	70
146	586	72637	14826	7	Flint	43	84
147	587	72645	14898	6	Green Bay	44	88
148	588	72655	14926	6	St. Cloud	44	94
149	589	72662	24090	7	Rapid City	44	103
150	590	72681	24131	7	Boise	44	116
151	591	72694	24232	6	Salem	45	123
152	592	72712	14607	6	Caribou	47	68
153	593	72722	04734	6	Maniwaki	46	76
154	594	72734	14847	6	Sault Ste. Marie	46	84
155	595	72747	14918	6	Intern'l Falls	49	93
156	596	72764	24011	6	Bismarck	47	101
157	597	72768	94008	6	Glasgow	48	107
158	598	72775	24143	6	Great Falls	48	111
159	599	72785	24157	6	Spokane	48	118
160	600	72793	24233	6	Seattle	46	122

Sta No.	Trav. No.	WMO No.	WABAN No.	Lat. Bd	Name	Lat.	Long.
161	601	72798	24240	6	Tatoosh Island	48N	125W
162	602	72807	14508	6	Argentia	47	54
163	603	72811	15613	5	Seven Islands	50	66
164	604	72815	14503	6	Stephenville	49	59
165	605	72816	15601	5	Goose Bay	53	60
166	606	72826	15703	5	Nitchequon	53	71
167	607	72836	15803	5	Moosonee	51	81
168	608	72848	15806	5	Trout Lake	54	90
169	609	72867	25004	4	The Pas	54	101
170	610	72879	25111	5	Edmonton	54	114
171	611	72896	25206	5	Prince George	54	123
172	612	72906	15605	4	Fort Chimo	58	68
173	613	72907	15704	4	Port Harrison	58	78
174	614	72909	16603	3	Frobisher Bay	64	69
175	615	72913	15901	4	Churchill	59	94
176	616	72915	16801	3	Coral Harbor	64	83
177	617	72917	18801	1	Eureka	80	86
178	618	72924	17901	1	Resolute	75	95
179	619	72926	16903	3	Baker Lake	64	96
180	620	72934	26102	3	Fort Smith	60	112
181	621	72938	26107	2	Coppermine	68	115
182	622	72945	25218	4	Fort Nelson	59	123
183	623	72957		2	Dease Lake	68	134
184	624	72964	26316	3	White Horse	61	135
185	625	72968	26317	2	Aklavik	68	135
186	626	74043	26202	2	Norman Wells	65	127
187	627	74051	27201	1	Sachs Harbor	72	125
188	628	74072	27101	1	Mould Bay	76	119
189	629	74074	27001	1	Isachsen	79	104
190	630	74081	16895	2	Hall Lake	69	81
191	631	74082	18601	1	Alert	83	62
192	632	74090	17601	2	Clyde	70	69
193	633	74109	25223	5	Port Hardy	51	127
194	634	74486	94789	7	New York	41	74
195	635	74794	12868	10	Cape Canaveral	28	81
196	636	76458	22009	11	Mazatlan	23	106
197	637	76644	12878	11	Merida	21	90
198	638	76679	11903	11	Tacubaya	19	99
199	639	76692	11904	11	Veracruz	19	96
200	640	78016	13601	9	Kindley (Berm)	32	65

Sta No.	Trav. No.	WMO No.	WABAN No.	Lat. Bd	Name	Lat.	Long
201	641	78063	12712	10	Grand Bahama Is	27N	78W
202	643	78089	12716	11	Bonefish Bay	24	75
203	644	78118	12714	11	Grand Turk Is	21	71
204	645	78325	12864	11	Havana	23	80
205	647	78367	11706	11	Guantanamo	20	75
206	648	78383	11813	11	Grand Cayman	19	81
207	649	78397	11715	11	Kingston	18	77
208	650	78467	11646	11	Sabanade Ma	19	69
209	652	78501	11807	11	Swan Island	17	84
210	653	78526	11641	11	San Juan	18	66
211	654	78806	10701	13	Balboa	09	80
212	655	78861		11	Antigua	17	62
213	656	78862	11647	12	Antigua	17	62
214	657	78866	11645	11	St. Martin	18	63
215	658	78897	11642	12	Le Raizet	16	62
216	659	78967	11621	12	Chaguaramas	11	61
217	660	78988	11643	12	Curacao	12	69
218	661	80001	11814	12	San Andres	13	82
219	663	91066	22701	10	Midway Island	28	177
220	664	91115	42401	10	Iwo Jima	25	141E
221	665	91131	42502	10	Marcus Island	24	154
222	666	91165	22536	11	Lihue	22	159W
223	667	91217	41415	12	Taguac	15	145E
224	669	91245	41606	11	Wake Island	19	167
225	670	91250	41601	12	Eniwetok Atoll	12	162,
226	671	91275	21603	11	Johnston Island	17	170W
227	672	91285	21504	11	Hilo	20	155
228	673	91334	40505	13	Truk	07	152E
229	674	91348	40504	13	Ponape	07	158
230	675	91366	40604	13	Kwajalein	09	168
231	676	91376	40710	13	Majuro	07	171
232	677	91408	40309	13	Koror	07	134
233	678	91413	40308	13	Yap	10	138
234	679	91700	60703	13	Canton Island	03S	172
235	680	98327	41207	12	Clark AFB	15N	121
236	682	99041		6	Weather Ship	45	16W
237	683	99052		5	Weather Ship	53	20
238	684	99061		4	Weather Ship	59	20
239	685	99063		3	Weather Ship	62	33
240	689	99223		10	Weather Ship	29	135E

Sta No.	Trav. No.	WMO No.	WABAN No.	Lat. Bd	Name	Lat.	Long.
241	693	99360		2	Weather Ship	66N	02E
242	694		03124	9	Ft. Huachucha	32	110W
243	695		03125	9	Yuma	33	115
244	696	02B		4	Weather Ship	57	51
245	697	03C		5	Weather Ship	53	36
246	698	04D		7	Weather Ship	44	41
247	699	05E		9	Weather Ship	35	48
248	700	17P	15045	5	Weather Ship	50	145
249	701	24N		10	Weather Ship	30	140
250	702	25V		9	Weather Ship	34	164E

Table A-2. Number of individual observations containing both wind and temperature as a function of latitude belt and level.

	100 mb	50 mb	30 mb	20 mb	15 mb	10 mb
52°N (V)	13,017	9,427	6,179	3,682	2,245	816
47°N (VI)	29,913	23,677	17,366	11,419	6,889	2,766
42°N (VII)	34,249	29,496	25,781	16,125	9,818	3,293
37°N (VIII)	29,040	22,852	19,260	12,364	7,178	3,494
32°N (IX)	30,332	23,467	20,677	14,272	9,367	4,369
28°N (X)	22,549	17,177	14,462	10,394	7,629	4,384
22°N (XI)	19,231	15,344	12,071	6,986	4,311	1,877
14°N (XII)	14,342	9,535	8,204	6,236	4,455	2,221
8°N (XIII)	7,214	4,684	3,446	2,407	1,791	1,105

Table A-3. List of stations (southern hemisphere).

WMO No.	WABAN No.	Lat. Bd	Name	Lat.	Long.
85442	52701	20°S	Antofagasta	23S	70W
61900	50101	8°S	Ascension Island	8	14
94335		20°S	Cloncurry	20	140E
96996		12°S	Cocos Island	12	96
94120		12°S	Darwin	12	131
82400	50301	3°S	Fernando Noronha	4	32W
84129	50701	3°S	Guayaquil	2	80
84631	51701	12°S	Lima	12	77
66160		8°S	Luanda	9	13E
68032		20°S	Maun	20	23
94312		20°S	Port Hedland	20	119
82898	50303	8°S	Recife	8	35W
83781	52402	20°S	Sao Paulo	23	46
94294		20°S	Townsville	19	146E
68110		20°S	Windhoek	22	17
80222	10717	8°N	Bogota	4N	74W

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BIOGRAPHICAL NOTE

The author was born in New York City on October 28, 1940. He graduated from Bayley-Ellard High School in Madison, N.J. in June, 1958. He received his undergraduate training at Webb Institute of Naval Architecture in Glen Cove, N.Y. His work experience as part of the program included an apprenticeship at the New York Naval Shipyard, a tour of duty on a merchant ship, and several periods of employment as an engineering aid at David Taylor Model Basin. He was awarded a B.S. degree in naval architecture and marine engineering in June, 1962.

While at M.I.T. he has been the receipient of a Ford Foundation fellowship for one year and a Fannie and John Hertz Engineering Scholarship Foundation fellowship for two years.

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