

# ATOLL RESEARCH BULLETIN

*Some aspects of the meteorology  
of the tropical Pacific  
viewed from an atoll*

by

*Ronald L. Lavoie*



Issued by

THE PACIFIC SCIENCE BOARD

National Academy of Sciences—National Research Council

Washington, D. C., U.S.A.

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It is of interest to note, historically, that much of the fundamental information on atolls of the Pacific was gathered by the U.S. Navy's South Pacific Exploring Expedition, over one hundred years ago, under the command of Captain Charles Wilkes. The continuing nature of such scientific interest by the Navy is shown by the support for the Pacific Science Board's research programs during the past fifteen years.

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ABSTRACT

A detailed climatological study of surface and upper-air data from Eniwetok Atoll was carried out in order to throw light upon various tropical oceanic problems including island influence, diurnal variations, and the nature of synoptic disturbances. Among results of the analysis are: 1) the atoll influence upon cloud or precipitation over the atoll itself is hardly detectable and probably insignificant; 2) the diurnal variation of the surface wind is almost exclusively the result of world-wide, atmospheric tidal motions, except for a possible speed increase of less than 0.1 kt near mid-day; 3) low cloudiness shows a diurnal variation whose range is less than 4% of total sky with maximum cover at 0700 local time; 4) precipitation occurrence shows a distinct early morning maximum; 5) in synoptic disturbances affecting the region, equatorward flow gives rise to a slight tendency for better than normal weather while poleward flow is associated with markedly wetter weather; 6) disturbances which make themselves felt primarily in the wind speed field appear to contribute significantly to bad weather periods at Eniwetok.

The problem of diurnal rainfall variation is elaborated by inclusion of data from a weather ship and several other atoll stations. Various hypotheses are examined and it is concluded that the variation in shower frequency is most likely the result of cumulative stabilization of the cloud layer by absorption of solar energy in the cloud tops by day. The variation in low cloud cover is felt to be due primarily to a diurnal variation in sensible heat transfer across the air-sea interface.

## Introduction

The renaissance which tropical meteorology has experienced since about 1940 has been channelled quite understandably in the direction of a better understanding and prediction of the life cycle of tropical cyclones. In spite of the tremendous significance of improvement in this area, there are meteorologists who will remind us that the important tropical problems demanding solution are not limited to the circulation of the tropical cyclone. In this group of meteorologists are the forecasters in those areas generally free from these destructive storms who must face the vexing problems provided by the "undisturbed" or weakly disturbed tropical atmosphere. Also there are those who feel that the improvement in our understanding and prediction of the tropical cyclone will be, or indeed may already be, stunted by the deficit of knowledge of the smaller scale or more normal processes of the tropical atmosphere. Finally, the student of the general circulation of the global atmosphere will remind us that the heat and moisture fed into the tropical atmosphere plays the role of prime mover for the large scale motion systems of ocean and air. Most of this driving energy and its day to day variations must be provided through processes apart from tropical cyclones.

The tropical latitudes are largely oceanic. The proper framework for building an understanding of the tropical atmosphere must rest upon a precise description of ocean-atmosphere interaction processes and accurate measurements of the atmosphere over the sea. But where does one gather such information? There have been no weather ships stationed

deep within the tropics, and oceanographic expeditions rarely stay in one place long enough to provide definitive description.

Islands are plentiful only in restricted regions of the tropical oceans, yet these provide us with stable observing platforms which have contributed much of our knowledge of tropical maritime weather. However, the tropical atmosphere is frequently in such delicate balance with the sea surface that even a rather small island might conceivably leave an indelible imprint upon local weather and our picture of the low level atmosphere. Accounting for these local influences is generally far from a simple task. Furthermore, how can a rational understanding be conceived until one learns what the variations would be without the intrusion of a land mass?

The primary objective of this investigation has been to scrutinize the standard meteorological observations from a coral atoll to determine: 1) the extent to which such a tiny island disturbs its atmospheric environment; 2) the character of diurnal variations over the sea; 3) the contributions which a detailed climatological analysis of such records might be able to make on the question of the nature of low-level disturbances in the central tropical Pacific.

When problems such as these have been dealt with satisfactorily it should be possible to make more intelligent use of island and ship observations in the tropics. From the applied standpoint such information would be very helpful in building a proper frame of reference for the evaluation of orographic and heat source effects on tropical islands.

## Part I: Eniwetok As an Observing Platform

It is, of course, improper to ask whether a small island has any effect on the overlying atmosphere. Even the smallest islet has some effect, even if the influence be limited to the lowest few centimeters of air. Furthermore one must not neglect the action of the sub-surface land mass as an obstacle to normal oceanic processes. This might alter the surface-water properties in the neighborhood of the island and transmit its influence to the atmosphere indirectly.

One must first select the variable and the location of its measurement and then ask whether the influence of the island makes a significant impression. Conjuring up experiments to answer such a question is an interesting pastime, but under the constraint of existing data the approaches are few and the answers are perhaps only suggestive.

Eniwetok atoll was selected for intensive study for several reasons. It is firstly representative of a moderate sized atoll far removed from other land and overlain by a persistent trade wind regime throughout the year. Secondly, the atoll has been the site of special intensive data collection on several occasions as the result of nuclear test programs in the area. Thirdly, the University of Hawaii has on file some 150,000 punched cards containing hourly surface observations, radiosonde and rawinsonde data for much of the period of record at Eniwetok. This file is a portion of the Pacific data library transferred to the University by the meteorological support arm of Joint Task Force Seven.

Eniwetok is an isolated atoll situated at the northwest edge of the Marshall Islands group. It is some 2500 statute miles west-southwest of

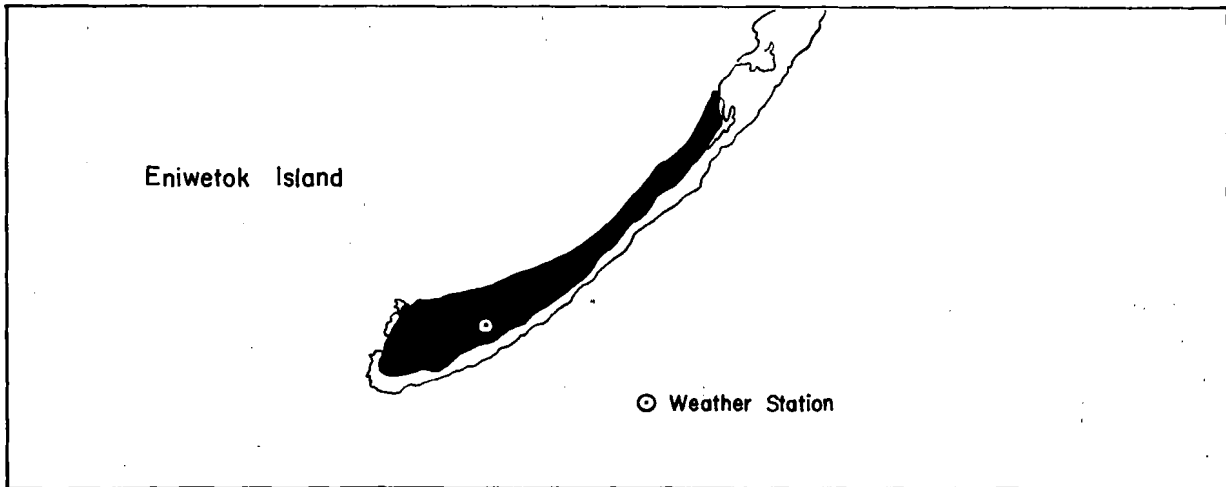
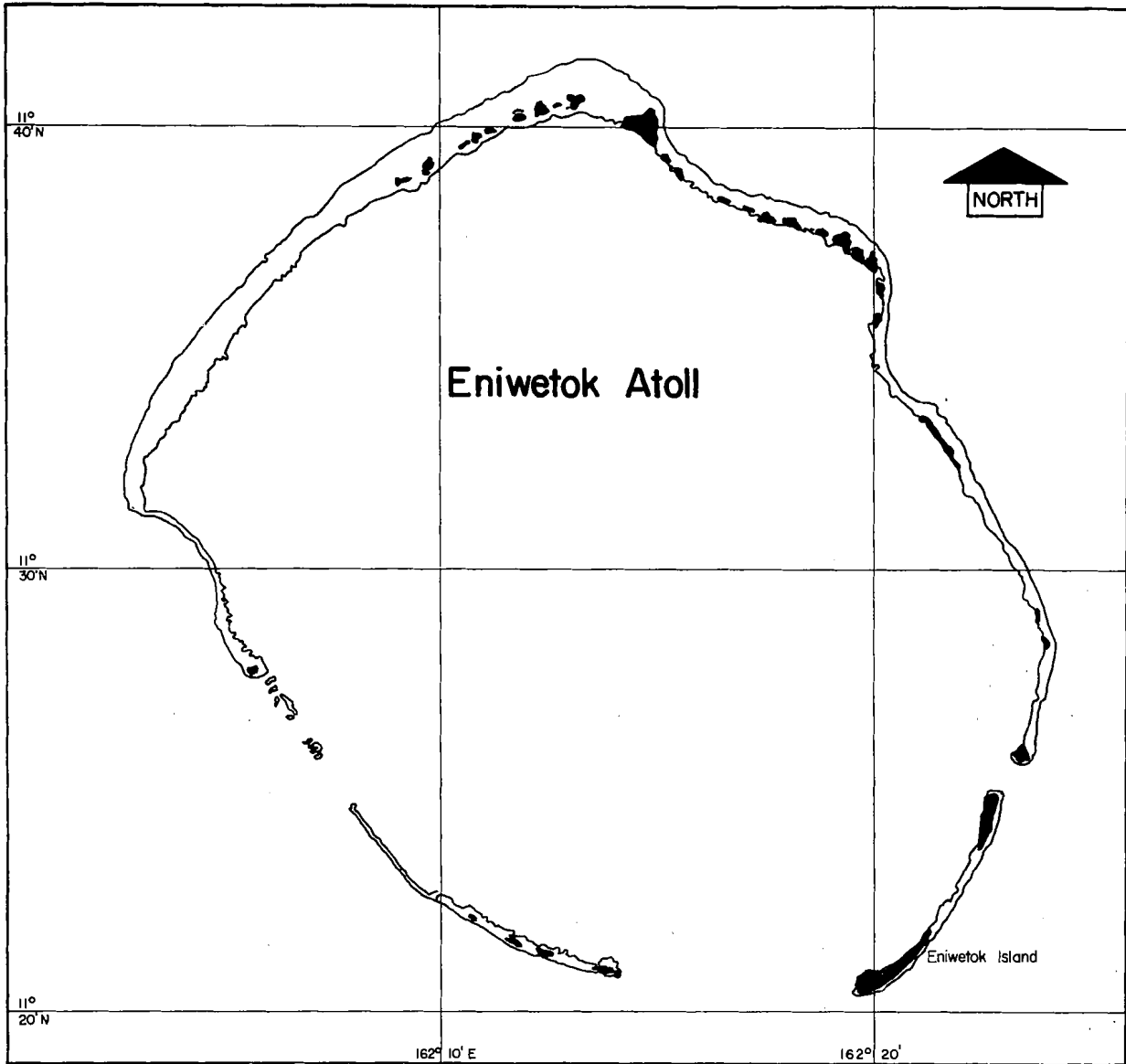


Fig 1. Outline map of Eniwetok Atoll. Dark areas depict dry land, clear outlines show outer margin of reef flats. Location of weather instruments is shown on enlarged chart of the main islet at bottom.

Honolulu at 11.4N, 162.3E. The atoll consists of a chain of about thirty small islets strung on coral reef around an oval lagoon 25 miles long by about 20 miles wide (Fig 1). Most of the islets are less than 13 ft high but some have coconut palms reaching to over 80 ft. The lagoon is quite deep, approximately 155 ft on the average. A detailed presentation of the geology and hydrography of the atoll is available in a report of the U. S. Geological Survey (Emery et al, 1954). A comprehensive summary of the topography, vegetation, and climate of Eniwetok is provided, along with original data from a brief, intensive microclimatic observational program, by Blumenstock and Rex (1960).

The only meteorological data of sufficient extent and homogeneity were collected near the main runway on Eniwetok Islet (Fig 1). It will be noted that this location is on the windward side of the atoll under prevailing east-northeast flow. The standard instruments are so situated that the air reaching them under typical trajectory probably traverses less than 400 meters of land surface.

With data from only one station on an atoll the evaluation of disturbances resulting from the atoll is difficult at best. An obvious approach is to determine the diurnal variation of various elements and to compare these with the expected trends resulting from the presence of a land surface. This will be done in the next section. Certainly the most important effects of the atoll would be the result of the contrasting heat transfer processes of land versus water and these should be exemplified by diurnal curves.

This approach is not completely satisfactory since there may remain some systematic influence present both day and night. One might expect,

for example, a different surface friction regime over the islets to have some effect on low-level winds. Also the lagoon water temperature may well be maintained at a slightly higher value than that of the surrounding free ocean where free circulation is unimpeded. The temperature difference is probably slight in this case with a deep lagoon and considerable tidal and wave exchange through channels and over inter-tidal reefs. Some scanty data are available (Blumenstock and Rex, 1960) which indicate that the lagoon surface water is usually less than 1<sup>o</sup>F warmer than the nearby ocean in August. A heat source of this magnitude seems unlikely to make itself felt in the cloud and precipitation regime.

An attempt to throw some light on this latter problem was made by a study of weather variations associated with the state of the oceanic tide. The reasoning here was as follows. Atolls generally possess a large reef area which is essentially exposed at low tide but submerged at high tide. In the case of Eniwetok the total dry land at high tide is only 2.5 mi<sup>2</sup> while at low tide exposed land (and tidal pools) covers some 34 mi<sup>2</sup> (Emory et al, 1954). This fourteen-fold increase is the result of a mean diurnal tidal range of 3.9 ft. It is reasonable to expect that the magnitude of the atoll heat source would be significantly increased therefore during those days when the low tide coincides with the hour of maximum heating. In addition to the order of magnitude difference in exposed land area there would be more shallow water and less lagoon water exchange at low tide.

Following this reasoning, the U. S. Coast and Geodetic Survey Tide Tables were examined for cases of maximum or minimum tide occurrence between the hours of 1200 and 1330 LMT for Eniwetok. Such an event will

be referred to as either a "high tide day" or a "low tide day". The dates of occurrence as well as the height of the tide and its range during the previous six hours were recorded for the period of record, June 1949 to February 1959.

The hourly surface reports were then examined for these dates and data were recorded for 0700, 1000, and 1300 LMT. The recorded observations included values of temperature, dew-point, low cloud amount and precipitation occurrence.

The mean increases of temperature and dew-point from 0700 to 1300 hrs were computed on low tide days and on high tide days and these were compared. It was hoped that this procedure would tend to reduce the possibility that any difference between the two sets of days could be due to chance synoptic influences. It might be expected that on days when the tide reached a minimum near 1300 hrs that the temperature (and possibly the dew-point also) would show greater increase from early morning to 1300 hrs. The analysis in fact revealed no significant difference between low and high tide days. The temperatures and dew-points showed mean differences of less than  $0.05^{\circ}\text{F}$  as a result of tide change. This result is not unexpected, however, since the tidal variation will have only a slight effect on the length of air parcel trajectory from water's edge to thermometer shelter (see Fig 1). The total heat transfer between the parcel and the surface must be insignificantly different from low to high tide situations.

This should not be the case with low cloud cover and precipitation, however. These quantities might be expected to respond to the area

effect of the heat source and to integrate the exposed reef increments over a large part of the atoll.

The analysis of precipitation could not be made satisfactorily due to the nature of the data. There were no hourly precipitation amounts available, only occurrence versus non-occurrence. Furthermore, the latter had to be taken from hourly "Airways" observations which report weather only during a 3 minute segment of the hour. Thus, in dealing with a small sample of days, the element of chance observations plays too large a role to allow statistical significance.

The amount of sky covered by low cloud for this 3-minute period is reported in tenths of total sky. This variable was carefully analyzed. In order to further limit the chance that a few synoptic disturbances might approach the atoll in the forenoon and give a random increase to one or the other sets of "tide days", values were compared at 1300 and 1000 LMT on each day. With tidal maxima or minima occurring near 1300 hrs on these days, 1000 hrs should represent the tidal node and these data should thus have little dependence upon the direction of tidal change. Those who have examined tidal records, however, will realize that not all "high tides" or "low tides" correspond to a significant change in elevation of the sea surface. This diluting factor was reduced here by ranking the cases according to the magnitude of the tidal deviation and selecting the top one third of each set of tide days so ranked for further analysis.

The change in low cloudiness from 1000 to 1300 LMT was then recorded for low tide days and for high tide days and values averaged for each of the 12 months. The comparison by month appeared advisable because of the

possible existence of a seasonal trend in the background diurnal variation of low cloud cover. Table 1 summarizes the results of this analysis which utilizes data from June 1949 to December 1958. Entries are in tenths of sky cover.

TABLE 1

Influence of Oceanic Tide on Low Cloudiness

Month	High Tide	Low Tide	$\Delta N_h$	No. of Pairs
	$N_h(13-10)$	$N_h(13-10)$	Col.1-Col.2	
Jan	-0.33	-0.47	0.14	15
Feb	-0.67	-0.63	-0.04	15
Mar	0.70	-0.17	1.23	15
Apr	0.07	-0.17	0.24	15
May	-0.13	1.23	-1.36	15
June	0.87	-0.47	1.34	15
July	0.53	0.13	0.40	15
Aug	-0.53	-0.50	-0.03	15
Sept	0.67	-0.27	0.94	15
Oct	0.10	0.07	0.03	15
Nov	0.17	-0.03	0.20	15
Dec	<u>-0.70</u>	<u>0.40</u>	<u>1.10</u>	<u>15</u>
TOTAL	2.05	-1.24	4.15	180

With the exception of May the values seem to indicate more cloud cover during afternoons with a high tide, although the mean cloud increase (May included) is only 0.35 tenths of sky. The result is contrary to expectations. A two-tailed test of significance utilizing the Student's "t" distribution shows no cause to accept this result as significantly

different from zero at the 95% level of confidence (this is largely due to the adverse May value which survived a thorough rechecking). With the variance demonstrated in these data a difference as large as 0.35 might be expected about 14% of the time in similar samples from a population of differences whose real mean is zero.

The only conclusion to be drawn is that the analysis was unable to show that an order of magnitude change in the exposed land surface of this atoll had any effect upon low-level cloud cover. This result might be used to argue that the atoll itself has no influence on low clouds. In defense of proponents of the alternate hypothesis, however, it should be pointed out that less than half of the sky viewed from the observing station lies over the atoll and that much of the extensive reef area lies across the lagoon (Fig 1). Furthermore it will be observed that the lagoon itself covers an area ( $360 \text{ mi}^2$ ) which is an order of magnitude larger than the land exposed at low tide. However, lagoon waters certainly provide a relatively feeble heat source when compared to land near mid-day.

The conclusion reached here is compatible with reports from meteorologists who have resided on atolls, i.e. that there appears to be no preferential location by quadrant, for trade cumulus build-ups. On the other hand it runs counter to many stories that claim experienced polynesian navigators could distinguish the presence of tiny islands in the distance by virtue of the cloud distribution. It must be stressed here that the foregoing analysis could only hope to uncover a fairly sizable effect observable from the atoll itself. It is possible that the disturbance in cloud organization is felt primarily downstream from the atoll. Some evidence in favor of this conclusion is found in the study of diurnal rainfall variation on Majuro Atoll (page 35).

Part II: Diurnal Variations.

General comments.

The punched card deck available for Eniwetok consisted of hourly "Airways" reports for the period from June 1949 to February 1959. Several months in this period were incomplete, however. There were scattered periods when only 3-hourly observations were recorded and other periods when one or more elements of the reports were missing irregularly. In compiling diurnal variations from this data file using the IBM 650, care was taken to assure accuracy and homogeneity of record. The machine was programmed to make elementary consistency checks on the data and to exclude those days when more than one observation was missing. Monthly mean hourly values were then computed for each month for each of the following: sea level pressure, wind speed, zonal and meridional wind components, temperature, dew-point, wet bulb temperature, relative humidity, low cloud amount and occurrence of precipitation. Each monthly mean value represents at least 200 days or about 7 years of homogeneous data.

In order to further elucidate the character of the diurnal variations, harmonic analyses were performed on the monthly and annual mean hourly values of pressure, wind, temperature and humidity. The analysis was carried to four harmonics for each variable according to standard procedures (e.g. Panofsky and Brier, 1958). The values of these harmonic components for the annual means are presented in Table 2. The representation for the magnitude of the variable  $x$  at local mean solar time  $t_i$  (in hours) is given by:

$$x = \bar{x} + C_1 \cos \left[ \frac{2\pi}{24} (t - t_1) \right] + C_2 \cos \left[ \frac{2\pi}{12} (t - t_2) \right] + C_3 \cos \left[ \frac{2\pi}{8} (t - t_3) \right] \\ + C_4 \cos \left[ \frac{2\pi}{6} (t - t_4) \right] + \text{H.H.} \dots \dots \dots (1)$$

where  $\bar{x}$  is the mean daily value of the variable,  $C_1$  to  $C_4$  are the amplitudes of the first to fourth harmonics respectively,  $t_1$  to  $t_4$  are the respective phase hours of each harmonic, and H.H. represents the contribution of the eight higher order harmonics which are needed for a complete representation. If the observed mean hourly values of  $x$  are plotted and compared with the plot of the single cosine curve given by the first harmonic in (1), the observed points will of course not all lie exactly on this curve. A measure of the contribution which each individual harmonic makes toward reconstructing the observed curve is given by forming the ratio of the variance of that harmonic curve to the variance of the mean diurnal curve. If this ratio is multiplied by 100 we have the percent of the observed variance which is "explained" by the given harmonic. This information is provided in the row headed "%V" in Table 2. The percent variance is additive so that when all 12 harmonics are utilized,  $\sum \%V_i = 100\%$ .

With all meteorological variables it is found that the diurnal and semi-diurnal harmonics together provide for most of the mean daily variation of the variable. Higher harmonics generally make increasingly smaller contributions and their physical significance become increasingly more questionable. They are more and more likely to be the reflection of random errors in the data.

TABLE 2

Annual Harmonic Components of Diurnal Variations of Pressure,  
Wind, Temperature and Humidity

	<u>Sea Level Pressure</u>	<u>E Wind Component</u>	<u>N Wind Component</u>	<u>Air Temperature</u>	<u>Dew Point</u>	<u>Wet Bulb</u>	<u>Relative Humidity</u>
C <sub>1</sub>	0.39 mbs	0.10 kts	0.07 kts	2.42 <sup>o</sup> F	0.51 <sup>o</sup> F	1.01 <sup>o</sup> F	4.6%
C <sub>2</sub>	1.03	0.20	0.18	0.84	0.26	0.39	1.4
C <sub>3</sub>	0.09	0.03	0.03	0.08	0.05	0.06	0.2
C <sub>4</sub>	0.02	0.02	0.03	0.18	0.05	0.08	0.3
t <sub>1</sub>	4.53 hrs	7.11 hrs	22.29 hrs	13.22 hrs	13.26 hrs	13.24 hrs	1.23 hrs
t <sub>2</sub>	10.04	9.99	6.76	0.38	11.47	11.98	6.18
t <sub>3</sub>	1.60	4.60	2.26	0.60	1.31	1.01	4.34
t <sub>4</sub>	4.28	1.72	23.24	3.48	3.20	3.34	0.58
%V <sub>1</sub>	12.6%	19.1%	13.3%	88.7%	73.2%	86.1%	90.9%
%V <sub>2</sub>	86.6	73.4	78.5	10.5	19.9	12.9	8.4
%V <sub>3</sub>	0.6	1.2	1.6	0.1	0.6	0.3	0.1
%V <sub>4</sub>	0.04	0.7	1.5	0.5	0.8	0.6	0.5
TOTAL %V	99.9	94.4	94.9	99.8	99.5	99.9	99.9
ANNUAL MEAN	1009.96 mbs	12.65 kts	3.73 kts	81.96 <sup>o</sup> F	74.46 <sup>o</sup> F	76.54 <sup>o</sup> F	78.45%

The phase hour,  $t_i$ , is to be interpreted as the first time on the 24 hour clock at which the  $i^{\text{th}}$  harmonic of that variable reaches a maximum. Obviously the  $i^{\text{th}}$  harmonic has  $i$  maxima and  $i$  minima over the 24-hour day. The hours are given in local mean solar time (LST minus 1.18 hrs). All subsequent references to time in this report will be local mean time (LMT) unless otherwise noted.

Consideration of the significance of the harmonics in Table 2 will be included in the following discussions of the diurnal variations of individual meteorological elements.

#### Sea level pressure

Both the descriptive and theoretical aspects of the mean diurnal variation of atmospheric pressure have been studied extensively for nearly 200 years. The most recent survey of this problem is probably given by Siebert (1961). It was early recognized that these variations were of larger amplitude in the tropics and were more easily studied at these latitudes where synoptic pressure disturbances were generally small and infrequent. Still there exist some unanswered questions or at least some weak hypotheses concerning several facets of the surface pressure oscillations. The results of an analysis of the mean diurnal pressure variation are presented here in some detail partly as a data contribution to the general descriptive problem of atmospheric tides but also as supporting information for the following discussion of diurnal wind variations.

The semi-diurnal component of the atmospheric tides has received the most concentrated attention and its description has achieved the most

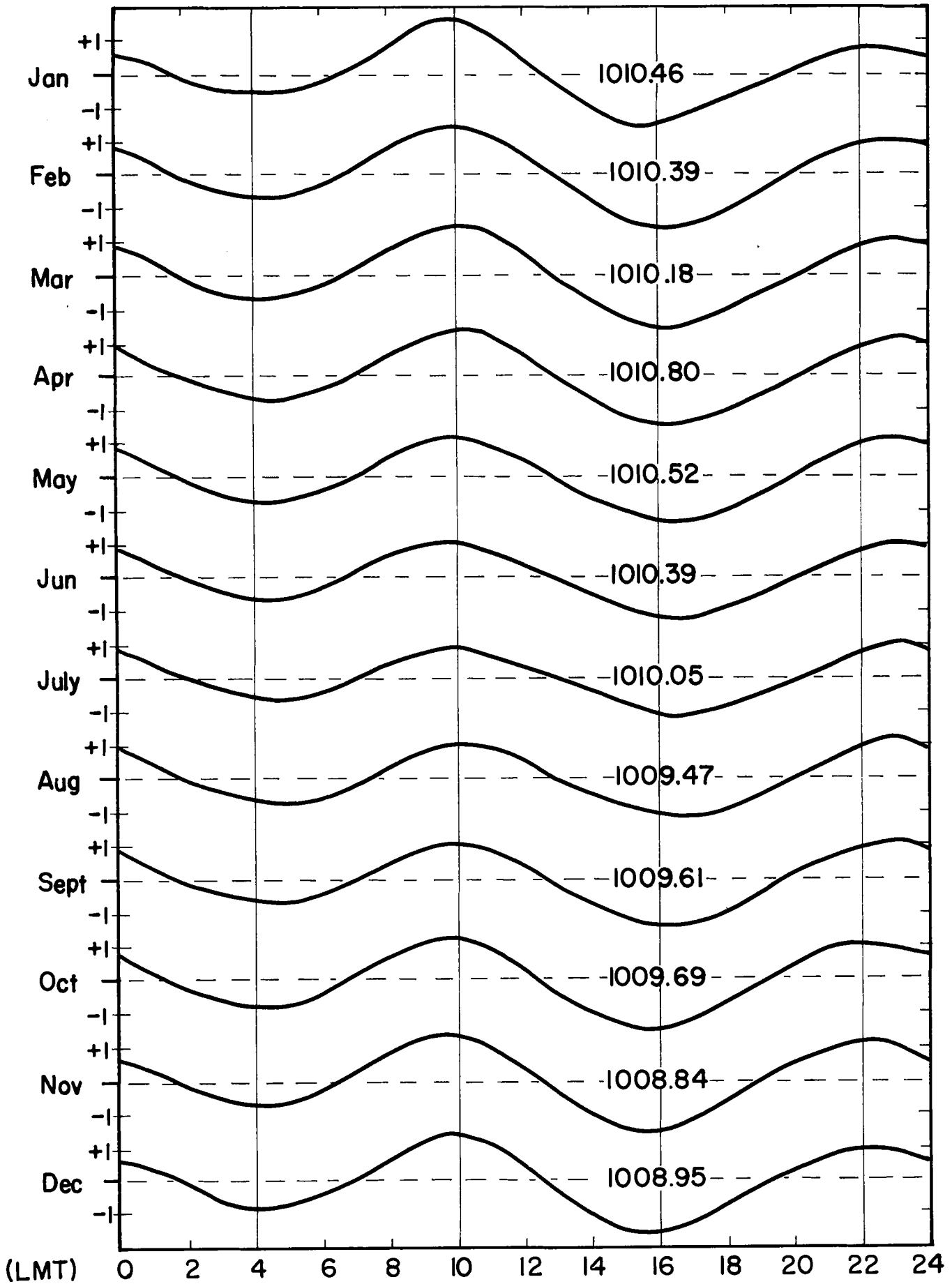


Fig 2. Mean monthly curves of diurnal variation of surface pressure at Eniwetok. Values are expressed as deviations from the monthly mean pressure in millibars. Time of day is local mean time.

universal agreement. The diurnal oscillation is consistently weaker, more susceptible to local influences (land-sea breeze, altitude, cloud cover, etc.), and less stable with respect to its seasonal fluctuations. Table 3 depicts the seasonal trend of the first three harmonic tidal components of pressure at Eniwetok, while Fig 2 presents the observed mean monthly curves.

TABLE 3

Monthly Mean 24-, 12-, and 8-hourly Pressure Oscillations.

	<u>C<sub>1</sub>(mbs)</u>	<u>t<sub>1</sub>(hrs)</u>	<u>C<sub>2</sub>(mbs)</u>	<u>t<sub>2</sub>(hrs)</u>	<u>C<sub>3</sub>(mbs)</u>	<u>t<sub>3</sub>(hrs)</u>
Jan	.45	5.19	1.08	9.88	.20	1.98
Feb	.47	5.11	1.12	10.10	.17	1.91
Mar	.51	5.27	1.13	10.13	.10	1.58
Apr	.49	5.27	1.10	10.14	.09	1.17
May	.36	4.71	1.02	10.16	.06	0.49
June	.36	4.47	.88	10.30	.07	7.88
July	.33	4.30	.86	10.46	.06	7.73
Aug	.33	3.65	.96	10.40	.04	0.78
Sept	.41	3.90	1.03	10.05	.05	1.01
Oct	.36	4.58	1.10	9.76	.11	1.89
Nov	.36	4.84	1.12	9.64	.14	1.94
Dec	.40	5.26	1.09	9.37	.17	2.13

The diurnal component of the atmospheric tides is a global phenomenon which is due primarily to the heating cycle of the solar day. The daily migration of this pressure wave around the earth is greatly distorted by local heat transfers but should be identifiable in its fundamental form at an atoll station in mid-ocean. No satisfactory

theoretical treatment of this oscillation has been given and no global description has been attempted to the author's knowledge.

The hypothetical picture of the global diurnal pressure wave given, for example, by Humphreys (1940, p 243) and Godske et al (1957, p 589) calls for a single progressive wave moving with the speed of the sun from east to west with maximum pressure observed at the coldest hours and minimum pressure in late afternoon. Observations on the U. S. mainland seem to support this hypothesis. However, recent work on Atlantic ship and island data (Haurwitz, 1955; Rosenthal and Baum, 1956; Harris et al, 1962) give hours of maximum pressure for the diurnal component ranging from about 1000 LST in the eastern to 2300 LST in the western Atlantic.

Chapman (1951, p 522) on the other hand states that the diurnal component is not a world wide oscillation, yet paradoxically indicates that where unaffected by local peculiarities this component would have a maximum at local noon.

The phase hour of the diurnal tide at Eniwetok is about 0500 and thus seems to agree with the deductions from the continental data. The amplitude of 0.4 mb is, however, about twice the value suggested by Godske et al (1957) as appropriate to "mid-Pacific pinpoint islands".

A hasty review of the literature on this subject leads one to suspect that the observations of the diurnal pressure wave may not be compatible with the hypothesis of a single progressive wave. There may well be a standing wave component which varies with universal time. There would seem to be a strong need for clarification of the nature of this fundamental atmospheric oscillation.

The seasonal variation of the diurnal component of the solar tide appears (from Table 3) to be quite regular. Largest amplitude occurs in late winter associated with latest phase hour. Minimum amplitude and earliest phase hour are found in late summer. This regularity in seasonal trend is unexpected in view of other published results (e.g. Rosenthal and Baum, 1956). Furthermore there seems to be a general consensus of opinion in the literature that the greatest amplitude should be found during the warmest months.

There is greater uniformity of results and hypotheses concerning the semi-diurnal component of the solar tide. Haurwitz (1956) gives the most recent global representation of this oscillation. The annual mean value of this component in Table 2 differs by less than .001 mbs in amplitude and 30 minutes in phase hour from values computed from his functional representation for this latitude and longitude. Haurwitz and Sepulveda (1957) have studied the seasonal variation of this component. The seasonal trend deduced by these authors is strongly supported by the results of Table 3. They hypothesize maximum amplitude in March and minimum in July, as observed. They further expect the earliest maximum to occur in winter and latest phase hour to be found in summer, again in perfect agreement with findings at Eniwetok.

The 8-hourly component of the atmospheric tide has received considerable attention, and a planetary representation of this oscillation has been given by Siebert (1957). The Eniwetok data show no significant disagreement with these results. One of the interesting facets of this oscillation is its near reversal of phase from summer to winter. This is a generally accepted feature of this harmonic and is demonstrated in Table 3.

### Surface Wind

The diurnal variation of the wind speed over most land areas is large, consistent, and easily rationalized. Winds tend to be strongest during hours of maximum solar warming and weakest during the cool early morning hours. The mechanism can be ascribed to the diurnal change of stability of the lowest air strata which affects the downward transfer of momentum from the stronger wind flows aloft. Diurnal speed variations over the sea appear to be the subject of some difference of opinion in the literature. Most textbooks of meteorology and climatology supplement the explanation of the continental situation with the statement that since diurnal stability changes are smaller over the oceans the amplitude of the speed variation is reduced. One is still led to expect some speed increase during the day. Bartrum (1957) for example, found a pronounced diurnal speed maximum at noon at Bermuda. The mean diurnal range was approximately 2 mph. On the other hand Brunt (1952) quotes Galle as reporting that at least during months May to October the maximum velocity of the SE trades of the Indian Ocean occurs during the night hours. Riehl et al (1951) in a study of the NE trades between California and Hawaii found an increase in wind speed from noon to midnight. In this latter study only two observation hours were compared, 0100 and 1300 LMT. The speeds were higher throughout the subcloud and cloud layers at 0100 hrs with differences at the surface of about 1.5 mph, reaching a maximum of nearly 4 mph near cloud base. No explanation for these observations was advanced. Riehl (1954) quotes work by Meinardus as well as that just mentioned to confirm the existence of diurnal wind variations at sea. Texts in oceanography, on the other hand, frequently give the impression

that the wind does not vary diurnally on the open ocean.

The mean hourly wind speed without regard to direction at Eniwetok is depicted in Fig 3 for a period of record of 2957 days. The striking similarity of this curve to that for the diurnal pressure variation (Fig 2) is immediately apparent. There can be no doubt that the atmospheric tides are the controlling factor for diurnal variations of surface wind on this atoll. The primary component here, as in the barometric oscillation, is the semi-diurnal.

The theory for the response of the wind to the semi-diurnal pressure wave has been given as early as 1910 by Gold and most recently examined in detail by Stolov (1955). According to this development the winds are antibaric, i.e. they blow anticyclonically around the two low pressure centers and cyclonically around the two high centers. The four pressure cells are strung, equally-spaced, around the equator. Thus at Eniwetok the tidal wind should be from the east as the high pressure cell passes at 1000 hrs, from the south at 1300 hrs, from the west at 1600 hrs as the pressure minimum arrives, from the north at 1900 hrs, etc. Since the prevailing wind is ENE with high steadiness, the tidal winds will account for the maximum speed near 1000 hrs (and 2200 hrs) and minimum near 1600 hrs (and 0400 hrs).

Godske et al (1957, p 590) argue that the winds produced by the diurnal pressure wave should also be antibaric, at least between  $30^{\circ}$  N and  $30^{\circ}$  S. This would explain, then, the close similarity between the pressure and wind curves; each component contributes in the same sense to the tidal winds.

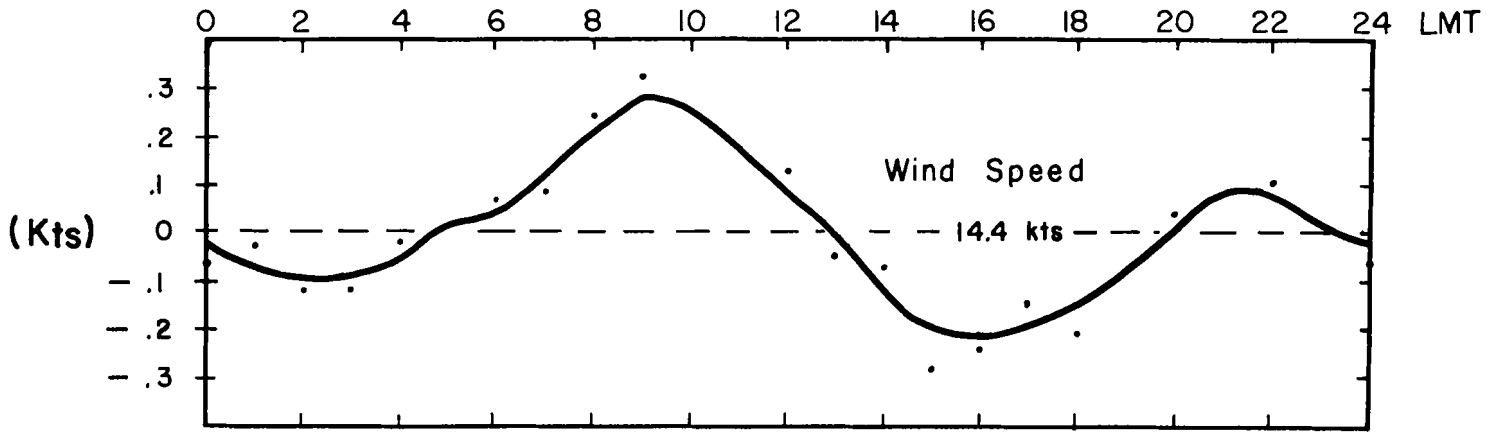


Fig 3. Mean diurnal variation of speed of the surface wind at Eniwetok. Curve shows 3-term running mean; dots are actual hourly values. Dashed line shows mean annual value.

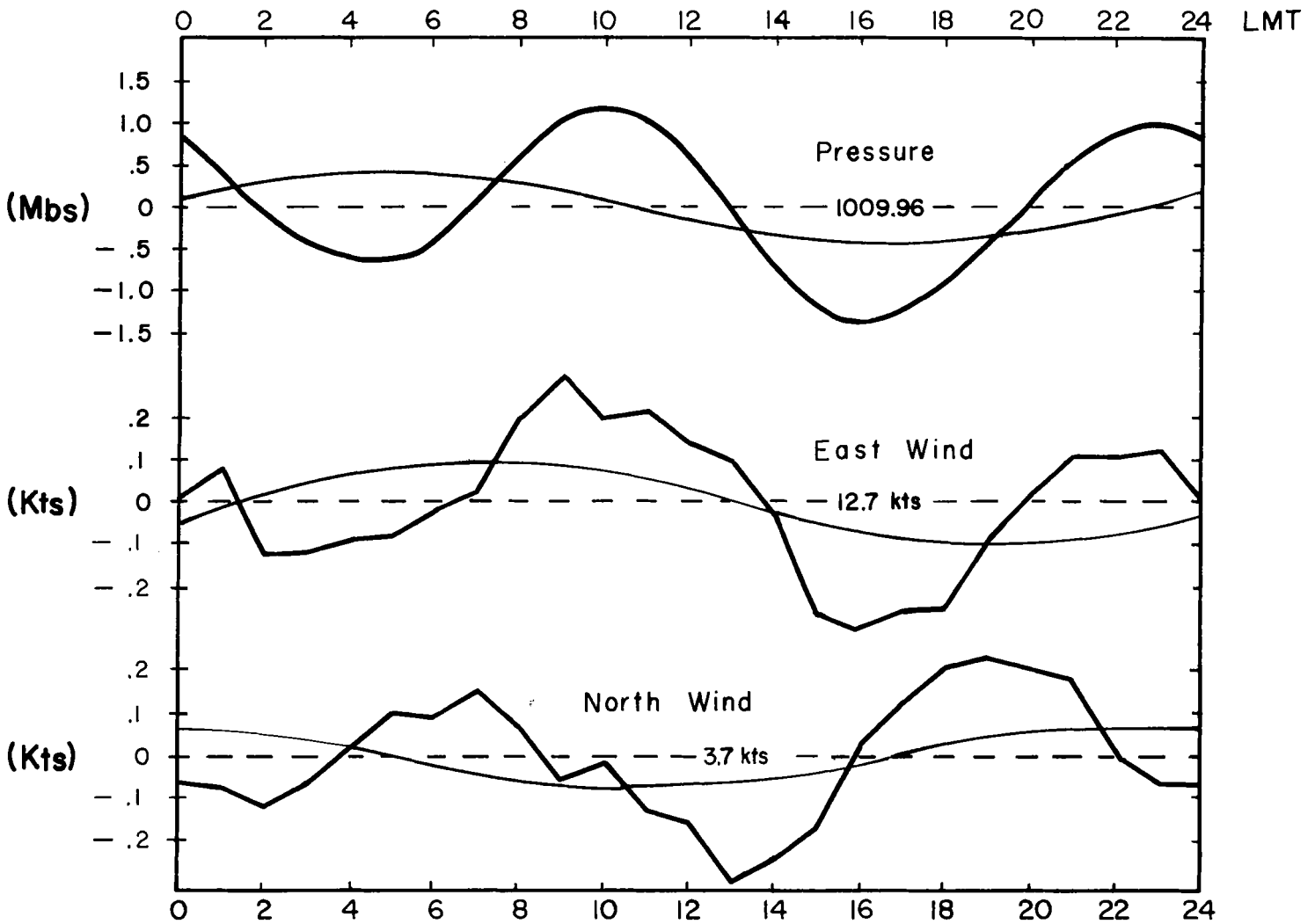


Fig 4. Mean diurnal variation of surface pressure and of East and North components of the surface wind at Eniwetok. Heavy solid lines connect actual hourly values. Thin solid curves are 1<sup>st</sup> harmonics of the diurnal variation.

In order to examine more critically the relationship between pressure and winds in the atmospheric tides problem, a harmonic analysis was made of the hourly mean wind vector. The data were not ideal for this purpose since wind direction was reported to only 16 points of the compass. This decreases the sensitivity of the data somewhat but the period of record is probably long enough (2715 days) to surmount this difficulty.

The diurnal curves of observed east (u) and north (v) components along with sea level pressure (P) are given in Fig 4 together with the 24-hour harmonics computed therefrom. It is of interest to compare both the 24- and 12-hour harmonics (from Table 2) with those expected from theory for the latitude and longitude of Eniwetok. This done in Table 4.

TABLE 4

Comparison of Observed and Predicted Air Tides.

	<u>Diurnal Component</u>		<u>Semi-diurnal Component</u>	
	Observed	Godske <u>et al</u>	Observed	Stolov
u (cm/sec)	$\pm 5.0$	unknown	$\pm 10.3$	$\pm 22.8$
v (cm/sec)	$\pm 3.5$	unknown	$\pm 9.3$	$\pm 10.6$
$t_u$ (hrs)	7.11	4.53	9.99	9.81
$t_v$ (hrs)	22.29	22.53	6.76	6.81
P (mbs)	$\pm 0.39$	$\pm 0.2$	$\pm 1.03$	$\pm 0.93$
$t_p$ (hrs)	4.53	$\approx 5$ a.m.	10.04	9.81

Considering first the semi-diurnal component it is seen that the agreement is very good with regard to phase but that the predicted amplitude of the tidal motion is too high. This may perhaps partly be due to frictional retardation which was not included in Stolov's development or to a faulty representation of the amplitude of the pressure wave at these

latitudes. Stolov utilized Simpson's formulation composed in 1918. The predicted value of 0.93 mbs pressure amplitude is based on this expression.

The diurnal component, as mentioned earlier is much more poorly understood. There is, however, sufficient agreement with the ideas expressed by Godske et al (1957) to lend strong support to the idea that this variation is due primarily, if not completely, to a global tidal phenomenon. If there exists any residual diurnal wind variation not due to air tides, it must indeed be small. Close examination of the first two columns of Table 4 will reveal, however, some argument for the reality of such a small diurnal effect. It will be noted that the phase hour of the v component follows expectation very closely while that for the u component is advanced by more than two hours. For the sake of argument we may suppose that the wind speed at the surface increases slightly toward mid-day as it does over land. Since winds are predominantly easterly this effect would be felt in the u component but hardly at all in the v. The effect would be essentially to superimpose a second daily cosine curve which would tend to pull the early morning maximum of u toward noon and increase its amplitude somewhat. This shift relative to the pressure wave is observed (see Fig 4). Certainly, the data do not warrant definitive conclusions from such minute effects. It does seem plausible though that the "local wind" variation, apart from air tides, is limited to a speed increase with a maximum near noon of amplitude about  $3 \text{ cm sec}^{-1}$  or about .07 kts. There is, of course, no way of knowing whether this, if realistic, is representative of the atoll only or of the open ocean. After consideration of stability variations in the lower air over the open ocean it becomes more reasonable to ascribe this effect to the atoll.

Before leaving the subject of atmospheric tides it might be of interest to consider briefly the seasonal trend of the wind variation to see how closely it agrees with that of pressure.

Fig 5 depicts the seasonal trend of both the amplitude and phase hour of the 1<sup>st</sup> harmonic of pressure and wind components. A very curious feature presents itself here. The phase hour of pressure varies slightly but regularly as noted earlier. The phase hours of the u and v components, however, show a peculiar tendency to move steadily around the clock (in opposite directions) with the seasons. The feature may quite likely be spurious. If it is real, the explanation for it escapes the writer. The amplitude variation does demonstrate that a certain degree of randomness is present.

Fig 6 gives a presentation similar to that of Fig 5 for the semi-diurnal harmonic. Much greater correspondence is demonstrated here between the three variables, and the theory of atmospheric tides is quite well verified.

#### Low cloudiness

The great difficulty of securing accurate observations of middle and high clouds is well known. The situation is not helped at all by the prevalence of opinion (particularly in the 1940's) that the tropical sky is usually covered by a veil of cirrostratus -- whether or not it can actually be seen. Furthermore, observing high cloud at night is notoriously difficult. For these reasons, taken together with the objective of assessing the atoll influence, only low cloud data are examined in this report.

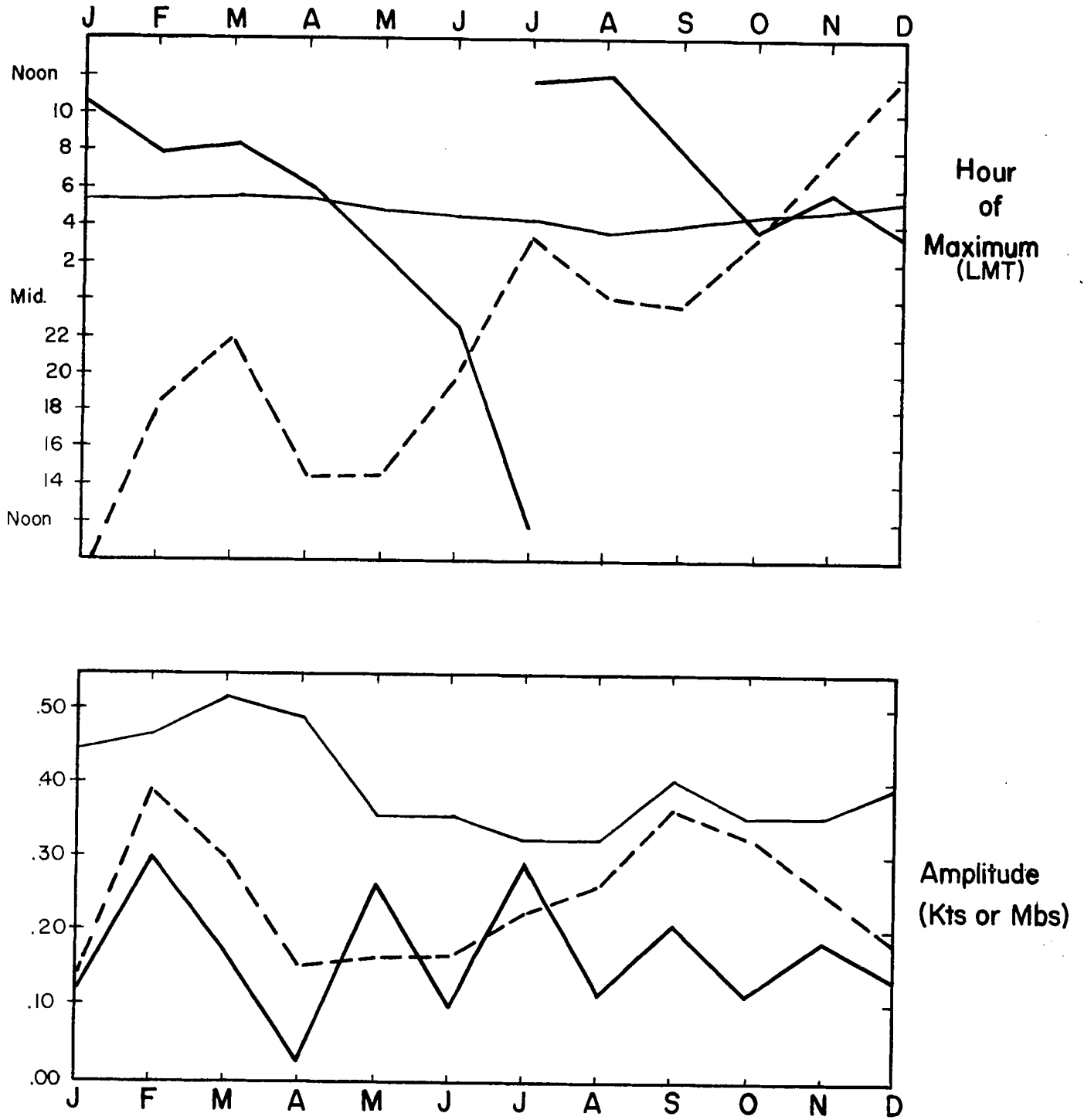


Fig 5. Seasonal trend of amplitude and phase hour of the 1<sup>st</sup> harmonic of pressure and wind components at Eniwetok. Heavy solid curves depict East wind component, dashed curves refer to North wind component, and thin solid curves refer to pressure.

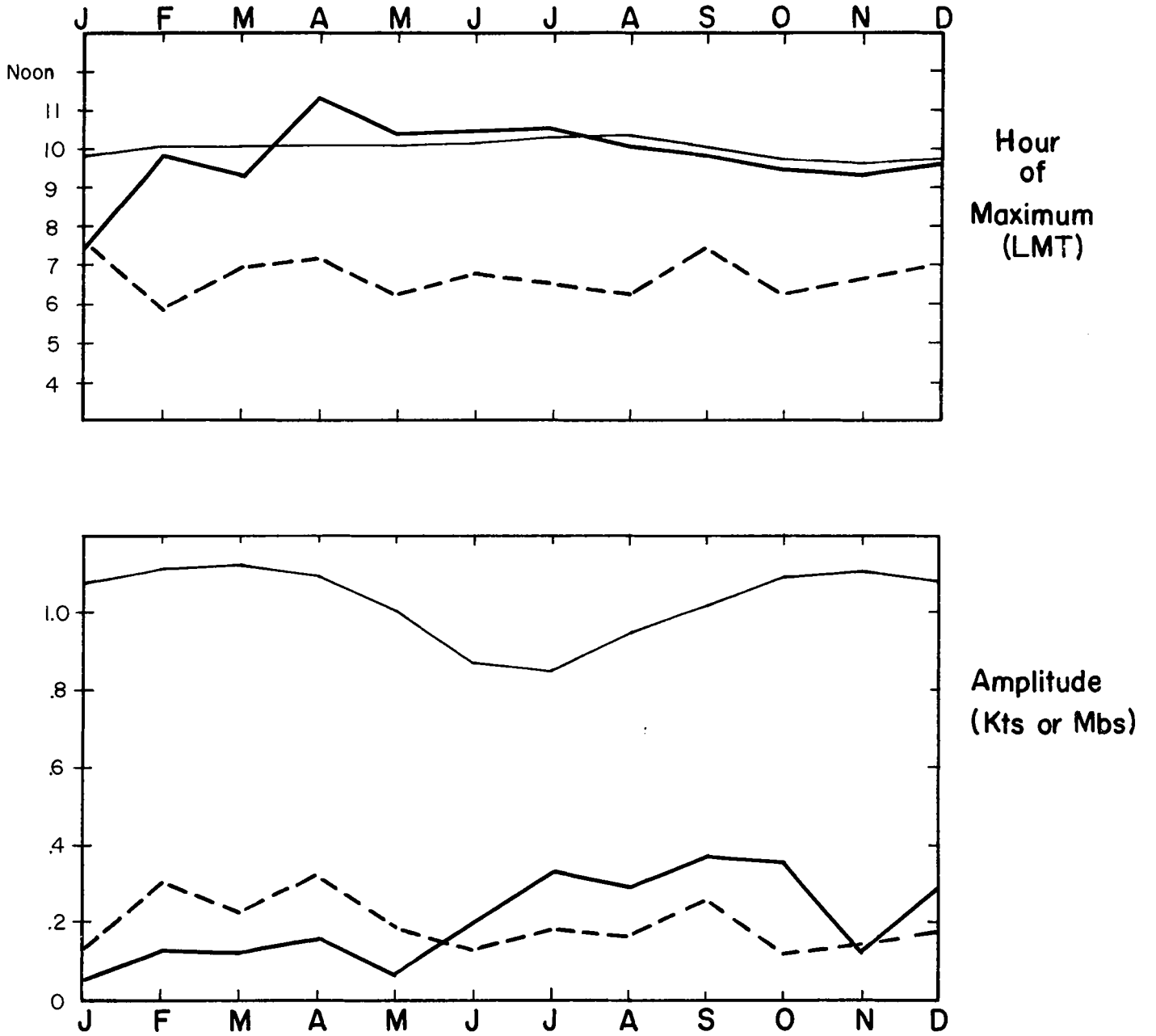


Fig 6. Same as Fig 5 except for the 2<sup>nd</sup> harmonic of pressure and wind.

The same precautions taken with the processing of other variables were followed with the cloud analysis, resulting in the selection of 2719 days of homogeneous data. No distinction was made between cloud types or cloud heights provided they were classified as stratus, stratocumulus, cumulus, or cumulonimbus. The overwhelming majority of low cloud types reported, however, was cumulus with bases at about 1800 ft. Whenever low clouds were reported at several levels the amount reported under a column titled "summary amount" was recorded. This latter value gives the decimal cover by all low cloud.

An analysis of the seasonal trend of the diurnal variation of cloudiness is given in Fig 7. The isopleths are based upon hourly values of low cloud amount expressed as percent deviation from the individual monthly mean. Monthly mean values vary little throughout the year, having a maximum of 3.2 tenths low cloud cover in May and minimum of 2.8 in September.

Fig 7 demonstrates some consistency in the diurnal variation throughout the year. Each month shows an increase in low cloud from midnight or earlier through the early morning hours. This feature is undoubtedly real. The only distinct seasonal trend is the tendency for an afternoon maximum in cloudiness to develop during the summer months, July through November. These months encompass the season of weakest trades, maximum sea surface temperatures, and greatest instability in the lower atmosphere. It is possible that the afternoon maximum during this season is brought about by the increased frequency of calms or light winds which result in greater heating of the surface waters during early afternoon. The atoll heat source would be expected to have its maximum effect under the same

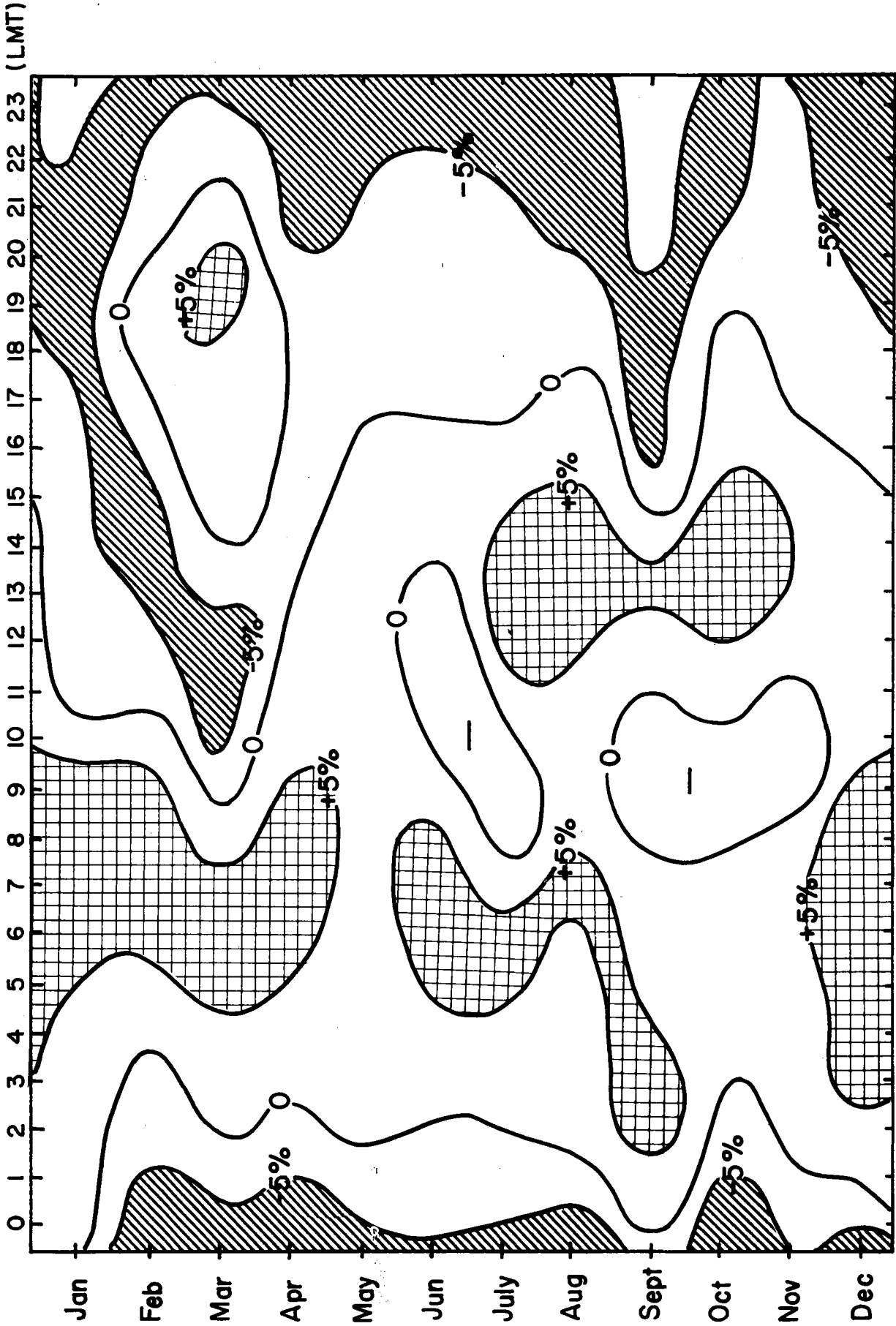


Fig 7. Diurnal variation of low cloud amount by months at Eniwetok expressed as percent deviation from the monthly mean. Grid pattern shows maxima, hatched areas are minima.

circumstances so that the two are difficult to separate. The results of the analysis reported in Part I are probably not sufficiently sensitive to uncover such slight variations of cloud amount. However, resumes of oceanographic cruises occasionally indicate a secondary maximum after noon (e.g. Garstang, 1958).

The main features of Fig 7 are quite evident in Fig 9 which depicts the annual mean diurnal variation curve. The daily range of cloudiness is seen to be quite small, 10-12% of the mean or 3.7% of total sky. The primary characteristic of the curve is undoubtedly the pronounced maximum at about 0700 and minimum near 2200 hrs. The small secondary maximum just after noon is the contribution of the weak trade season discussed in the previous paragraph.

There appears to be general agreement in the literature concerning the tendency toward maximum low cloudiness over oceans during the early morning hours although detailed studies of this point are very few. Riehl (1947) investigated data for 920 observing days from three ship stations near 35°N in the Atlantic Ocean. Using 3-hourly observations only, he found the diurnal variation of low cloud to be similar at each ship with a maximum at 0600 hrs (3.4 tenths) and minimum at 2100 hrs (3.2 tenths). These results show remarkable agreement with those reported here.

These results may be said to support the conclusions from Part I concerning the absence of a significant influence of the atoll upon low cloudiness. If the diurnal variation of cloud over the atoll differs negligibly from the variation observed at a ship station, one is led to believe that the influences of the two observing platforms differ

negligibly. Again, caution must be taken to admit the possibility of a systematic influence which does not react to the daily heating cycle. It seems clear, however, that textbook statements on the subject such as (Riehl, 1954, p 111): "Even atolls have a definite effect on the local cloud structure and its diurnal course", do not apply at Eniwetok.

If there can be said to be some consensus of opinion concerning the character of the diurnal variation of low cloud over the sea, there is yet little evidence of agreement on its cause. A discussion of hypotheses will be postponed until the nature of the hourly trend of precipitation is explored. It seems reasonable to consider the explanation of these two features in tandem.

#### Precipitation

The reality of the oft-reported nocturnal rainfall maximum over the ocean has never been fully investigated. The objection raised against most published records which demonstrate the existence of such a feature is that possible coastal or island influences have not been excluded. Certainly, most of the long term data on oceanic precipitation has been gathered on continental coastlines or on rather large mountainous islands. In either case it is difficult to evaluate the effects of land-sea breeze regimes, orography, and mountain-valley winds upon the "undisturbed" precipitation pattern. Nevertheless, it is common practice among climatologists to distinguish between maritime and continental precipitation regimes according to whether the diurnal maximum occurs by night or by day. The question remains, however: is the "maritime regime" representative of open ocean or only of coasts? Shipboard observations of precipitation are generally confined to occurrence versus non-occurrence.

(LMT)

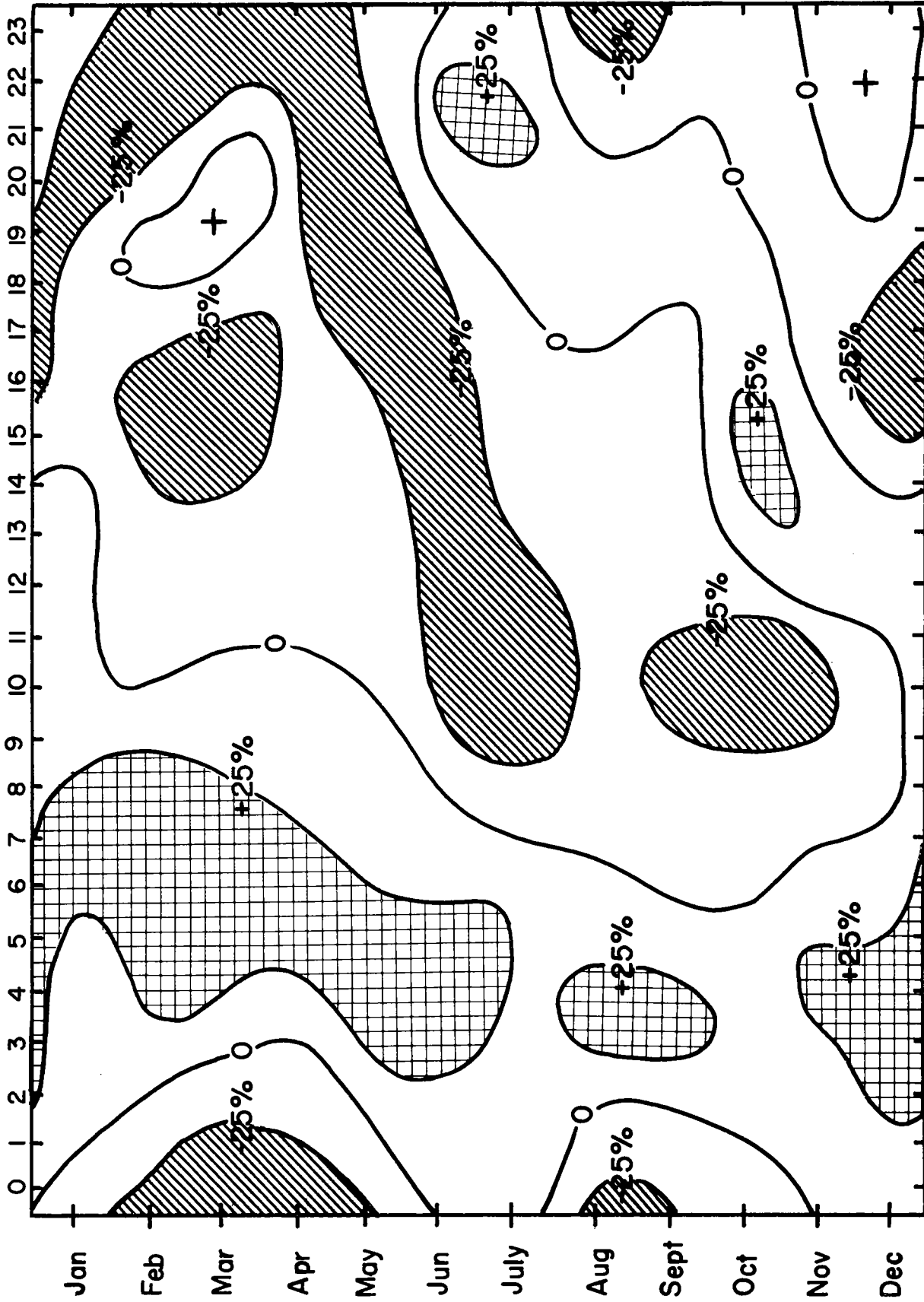


Fig 8. Same as Fig 7 but referring to precipitation occurrence.

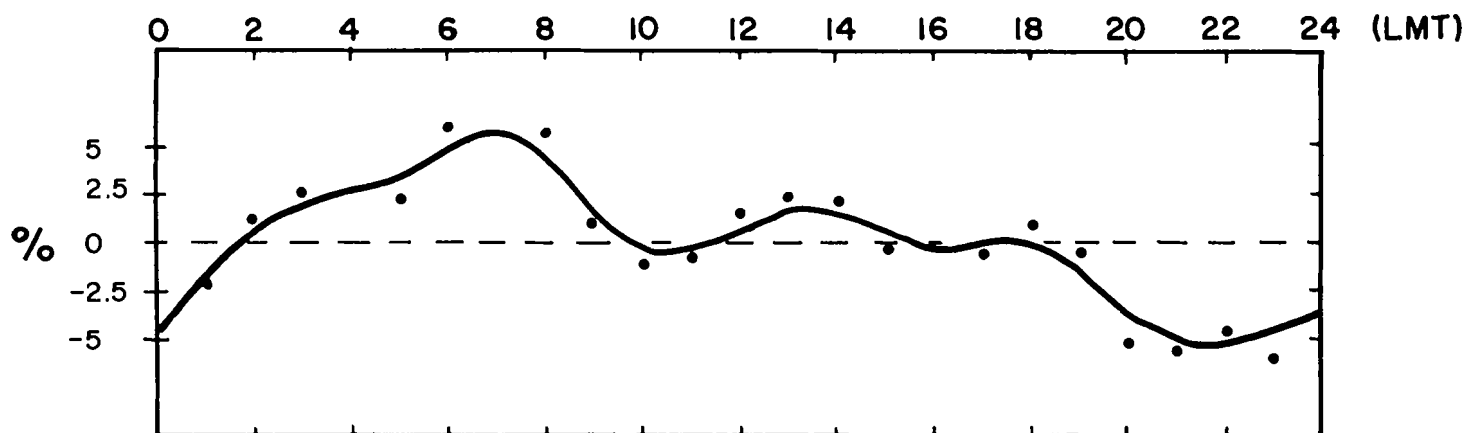


Fig 9. Annual mean diurnal variation of low cloud amount at Eniwetok. Deviations are in percent of annual mean (2.93 tenths). Solid curve is 3-term running mean of hourly values (heavy dots).

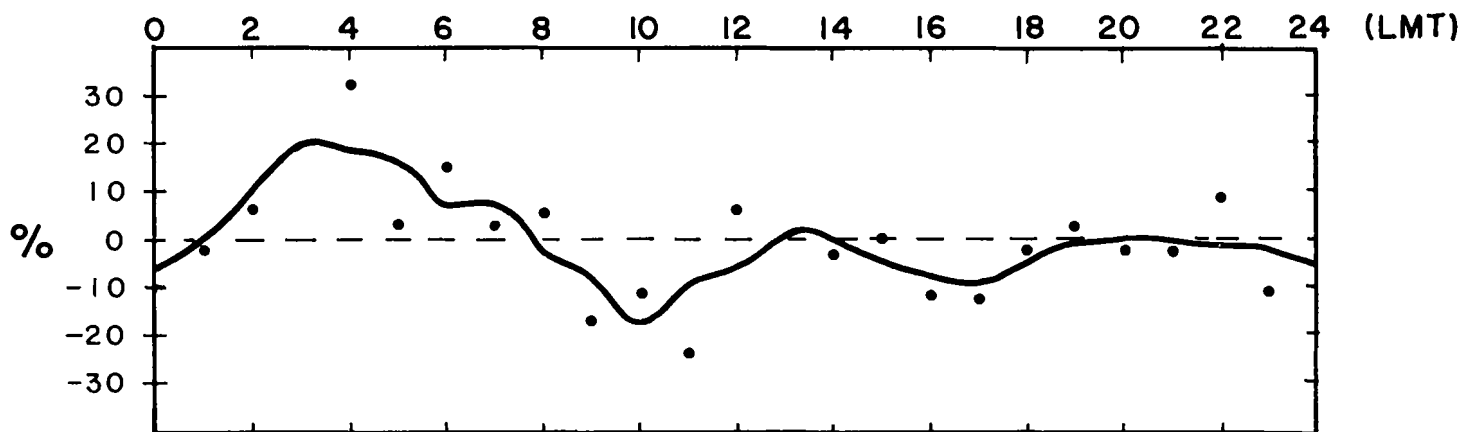


Fig 10. Same as Fig 9 but referring to precipitation occurrence at Eniwetok.

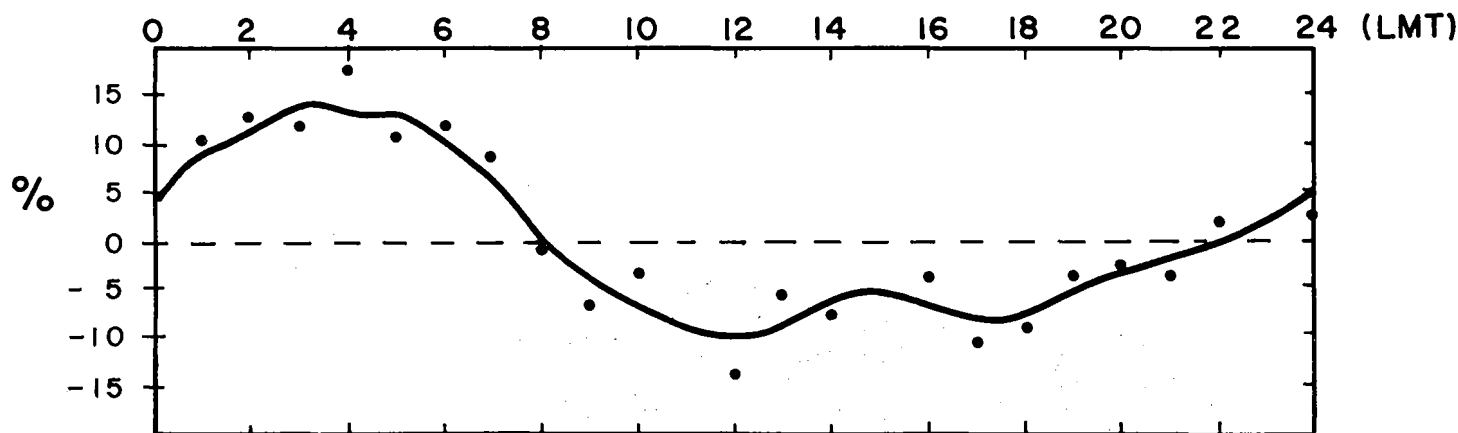


Fig 11. Same as Fig 10 but composited from precipitation occurrence at Wake, Kwajalein, Majuro, Johnston, Marcus, Canton, Christmas, Fanning and Palmyra Islands.

Furthermore these are generally limited to the 6- or 3-hourly synoptic reports. Still, a wealth of data exists in this area which has never been tabulated to this author's knowledge.

The hourly airways observations available for Eniwetok are meant to be a valid characterization of the weather between 2 and 5 minutes before the hour. Thus a shower which ends just before observation time is not recorded in the weather symbols. This effects a drastic reduction in the actual frequency of rain occurrence and reduces the size of the data sample considerably from the point of view of determining hourly variations. Rain amounts were totaled over 6-hourly periods, hence these are of only limited use in the characterization of diurnal variation.

Fig 8 gives a representation of the seasonal trend in the diurnal variation of precipitation occurrence at Eniwetok. Nearly 9 years of hourly totals were expressed as percent deviations from the monthly mean for the purpose of this analysis. From the foregoing discussion of the data it is to be expected that considerable randomness was evident in the actual values. These were first smoothed by a 3 term running mean process before analysis.

Some meaningful pattern is evident in Fig 8. There is an unmistakable nocturnal maximum in all months although it exists as only a secondary maximum in October (the wettest month). Definite seasonal trends are difficult to establish, however. In view of their tentative nature, seasonal differences will not be elaborated further. There would, indeed, appear to be justification for combining all months to reduce meaningless fluctuations.

Fig 10 summarizes the annual mean variation. The diurnal range of

hourly rain occurrence is seen to have a range of about 40% of the mean with a maximum between 0300 and 0400 hrs and a minimum near 1000 hrs. The secondary maximum between 1300 and 1400 hrs is largely a fortuitous feature as can be shown from Fig 8. In fact a broad minimum centered in the early afternoon appears to be a reasonable interpretation of Fig 8.

There seems to be little doubt that a significant early morning maximum in precipitation exists at Eniwetok. However, in view of the random fluctuations in the diurnal curve (Fig 10), certification of this point by examination of records from other atolls was felt to be desirable. A copy of Navy Job #3606 (1960) was secured from the National Weather Records Center for this purpose. This data summary presents an hourly tabulation of rainfall occurrences by month for several Pacific stations. This summary included the following atolls: Wake, Midway, Kwajalein, Majuro, Johnston, Marcus, Canton, Christmas, Fanning, and Palmyra. The periods of record varied from about 1 to 13 years, and the frequency of hourly observations was far from uniform. Still when the individual station records were normalized, all but one, Majuro, showed a definite nocturnal maximum falling between 0200 and 0600 hrs. A composite diurnal curve of all of these atolls is given in Fig 11 where the time scale refers to zone time for the nearest standard meridian. The composite was constructed by simply adding corresponding hourly total rain frequencies (after normalizing each to a representative observing frequency) and then computing deviations from the grand mean. Thus each station was essentially weighted according to its length of record. The curve represents the hourly distribution of a total of 35,700 individual observations of rain on the ten atolls.

The diurnal variation appears, from Fig 11, to be rather small. However, it must be kept in mind that most of the rainfall on these tiny islands is the consequence of the passage of synoptic disturbances. Certainly the convergence patterns associated with these must frequently overpower any tendency for diurnal regulation by the oceanic regime. Thus the diurnal range of about 25% (of the mean) would seem to be indicative of a real effect which demands explanation.

Nevertheless, it is troubling to find exceptions to this general diurnal trend and this may have deterred researchers from attempting a definitive explanation of the phenomenon. Ramage (1952), for example, found an insignificant and confused diurnal pattern on several island stations up to a few hundred miles off the Asiatic continent. At the same time most coastal stations showed a distinct early morning maximum. Of the atolls composited in Fig 11 Majuro is strikingly dissimilar in its diurnal march of rain occurrence. This atoll resembles Eniwetok in its physical features and in the location of its observing station relative to the lagoon and prevailing winds. It differs from Eniwetok, however, in its proximity to another atoll, Arno, some 10 miles to the east. Majuro is also situated in the Marshall Islands some 600 nautical miles ESE of Eniwetok at about  $7^{\circ}\text{N}$ ,  $171.5^{\circ}\text{E}$ . Fig 12 gives the diurnal curve of precipitation occurrence at Majuro tabulated from the station's Local Climatological Data (April 1958-August 1961) published by the U. S. Weather Bureau. These data are superior to those tabulated from Airways reports since they reveal the number of hours during which rain was observed at any time during the hour. The lower curve in Fig 12 shows a peculiar trend with the typical sharp increase in rain frequency from midnight to 0300 hrs followed by steady values throughout the

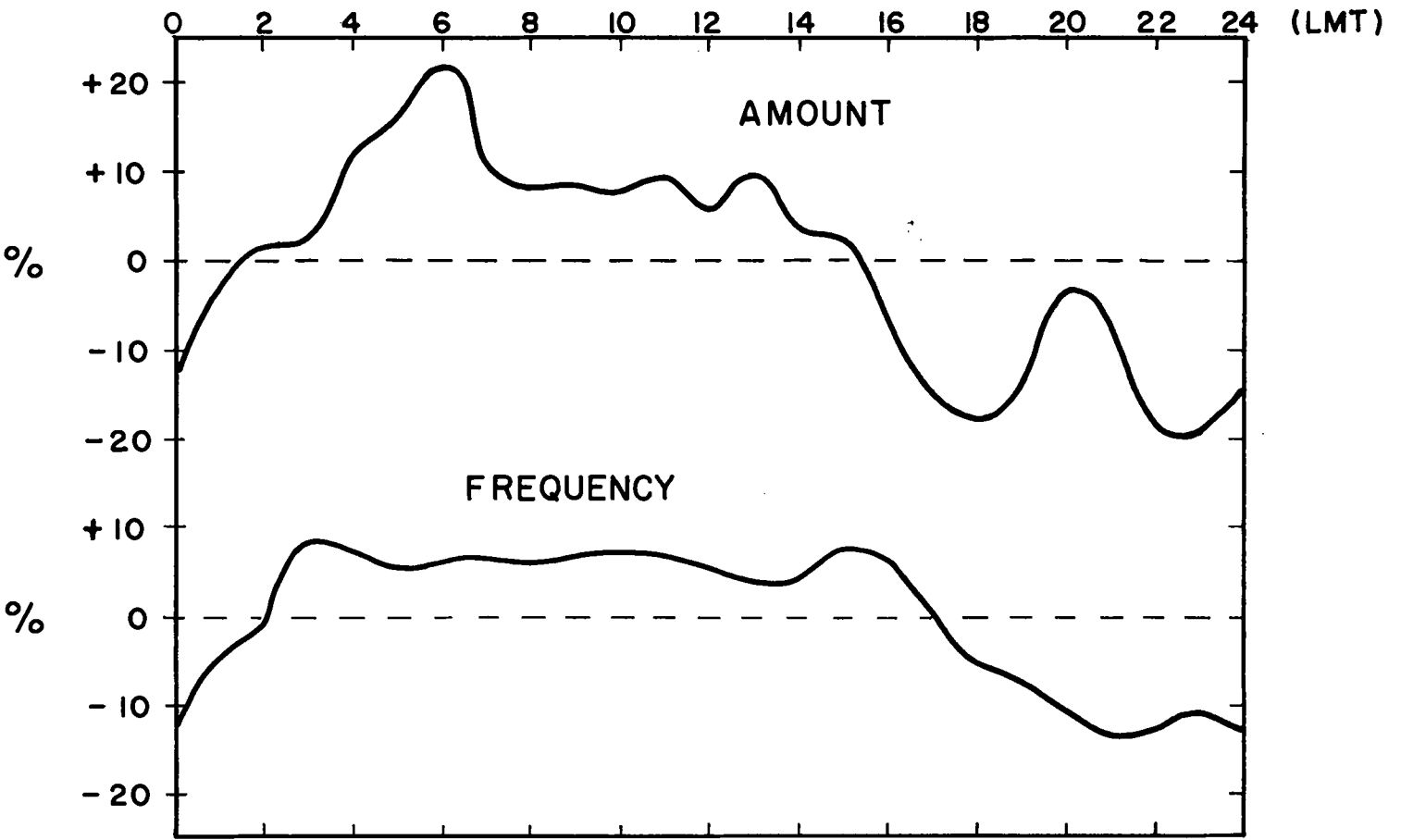


Fig 12. Annual mean diurnal variation of precipitation amount and occurrence at Majuro Atoll. Curves are 3-term running means of hourly values and describe percent deviation from annual mean.

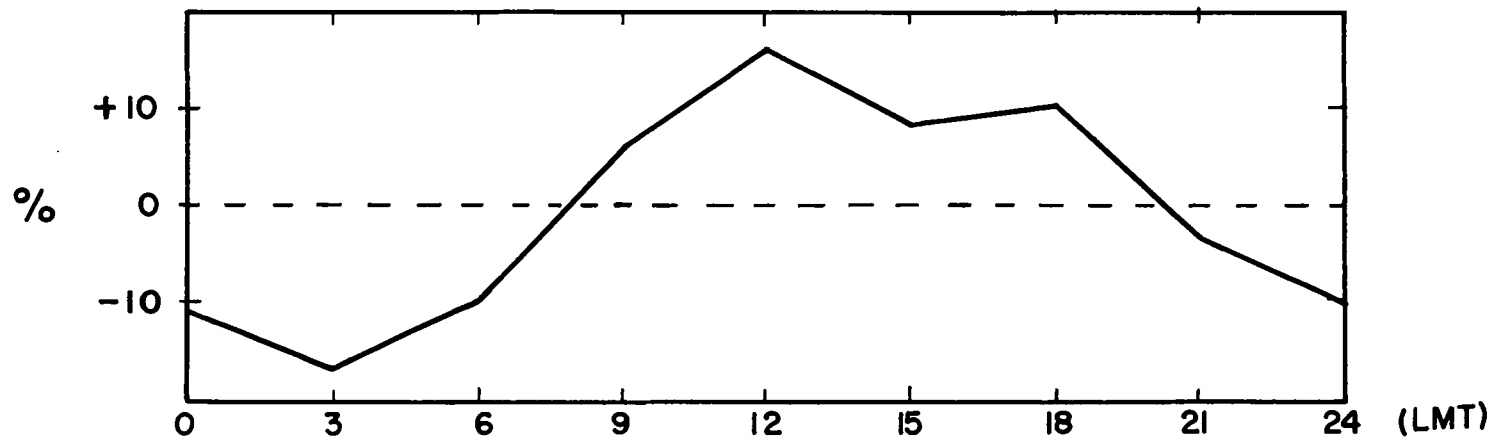


Fig 13. Annual mean diurnal variation of precipitation occurrence at Ship N (30°N, 140°W) taken from 3-hourly observations.

morning and until mid-afternoon. The upper curve shows the trend of hourly amounts taken by recording rainguage and extracted from the same source and time period as the first curve. An early morning maximum is evident here indicating that showers, if not most numerous, are at least most intense at about 0500-0600 hrs. It appears, then, that the "maritime influence" predominates here, but it may be contaminated by a downwind effect from Arno Atoll which has substantial land area.

A final piece of contradictory information will be included for the sake of completeness. Fig 13 depicts the diurnal rain occurrence trend at Ocean Station Vessel November positioned between Hawaii and California at  $30^{\circ}\text{N}$ ,  $140^{\circ}\text{W}$ . The period of record is January 1952 to June 1958. Data are from 3-hourly observations tabulated from Job #3606 (1960). Fig 13 indicates a diurnal trend which is about  $180^{\circ}$  out of phase with that quoted for atoll stations. No explanation for this phase reversal will be attempted here, but it may be compatible with observations of minimum height of the trade wind inversion during the night in this region. An explanation for such a diurnal variation of the strong inversion in the northeast trades at this location has been offered by Neiberger (1958). It was hypothesized that a standing wave is produced in the inversion layer by the forcing action of the sea breeze divergence at the coast. The height of the inversion base has a fundamental control on the development of precipitation at this station; a forced oscillation of this layer could undoubtedly exert primary control over shower frequency.

In summarizing the observations of diurnal variation of precipitation over the open ocean it is safe to say that the great preponderance of data indicates the reality of a late nocturnal maximum. It must be admitted

that exceptions can be found even on tiny islands or ship stations. However, it is likely that these can be successfully treated as exceptions, and other over-riding considerations may be found. From what has been deduced about the effect of atolls upon diurnal variations, it must be concluded that the phenomenon is at least representative of the mid-Pacific oceanic regions within the tropics. In view of this, it would be desirable to consider possible hypotheses in the light of the picture being constructed here from Eniwetok data.

The fundamental cause of a diurnal variation must be the sun, but the immediate causal agent could take several forms. Atmospheric tidal fluctuations appear quite incapable of bringing to bear an intensity of convergence greater than about  $5 \times 10^{-8} \text{ sec}^{-1}$ . At low cloud levels this could not induce vertical motions greater than  $10^{-2} \text{ cm sec}^{-1}$ . These must be considered negligible. Furthermore it has been noted earlier that the oscillation in the wind field is primarily semi-diurnal in character.

If the nocturnal maximum cannot be ascribed to a diurnal variation in convergence, we must look to the convectional-precipitation mechanism itself for a clue. Let us examine, for the sake of completeness, the variables which might possibly influence the precipitation potential of a cumulus cloud in the tropics. We have:

1. spectrum of condensation nuclei
2. cloud base temperature
3. updraft speed
4. turbulence spectrum
5. cloud depth
6. vertical shear of the wind

7. humidity of the cloud free air
8. width of cloud thermals
9. lifetime of the cloud
10. electrical properties of the cloud.

Information is lacking for a detailed consideration of each of these factors. However, we have no reason to suspect a diurnal variation in factors 1, 6, and 8. Factor 2 appears to be insignificant since there is no detectable diurnal variation in the reported heights of cumulus cloud bases, and free air temperature variations are small. Factors 3 and 4 depend largely on vertical stability in the convective layer. The diurnal trend of this variable should certainly be considered. Factor 5 will also be affected by the foregoing, particularly with regard to the height of the inversion or stable layer which limits development. Factor 7 undoubtedly has a diurnal variation though it is more likely to be an effect in this case than a cause. Factor 9 quite likely varies diurnally, but here also this could result from individual clouds being better developed at night and therefore requiring longer to decay. It is not a fundamental consideration here. Factor 10 is difficult to assess. However, it is well known that most atmospheric electricity parameters undergo a diurnal variation which, over the oceans, follows universal, not local time.

From cursory examination of the problem, then, it seems that a variation in vertical stability below or within the cloud layer and/or a variation in height of the inversion (or stable layer) must be suspected. We have ruled out as highly unlikely the possibility that the micro-physics of night-time clouds differs significantly from their daytime counterparts. Since individual clouds are short-lived, the colloidal stability of new

cloud matter would have to change in some pregressive fashion through the night to account for the rainfall trend. Likewise day or night will have a negligible effect upon the evaporation of raindrops between cloud and ground.

In order to examine the effect of thermal stratification on maritime cumulus we must first recognize an important distinguishing feature of these clouds, i.e. the absence of low-level convective "roots". Most observational evidence points toward an origin very near cloud base for most maritime cumulus updrafts. Thus the instability from which these clouds are born does not manifest itself in the surface layers as much as within the cloud layer itself. In this view, the heat supplied by the ocean surface (in the trades at least) is mixed upward by small scale eddies rather than organized thermals. This of course will not be the case if the sea is "sufficiently" warmer than the air. The exact stimulus for formation of such a cloud is an unsettled question but probably arises from either organized or random internal waves or turbulence at these levels which is able to cause sufficient vertical displacements to result in condensation. The conditional instability within this layer will allow such a cloud to grow until it reaches a stabilizing layer. One might expect, then, the variation in stability within and at the top of the cloud layer to have a controlling influence on cloud depth and also on rainfall. Let us for the moment concentrate our attention upon this 1 to 2 km layer.

Under trade wind conditions at the longitude of Eniwetok there seldom exists a pronounced inversion. The trade wind cumulus is usually capped instead by a layer of relative stability which is associated with an

increased lapse of mixing ratio with height. With clear air then, the the calculation of the divergence of flux of terrestrial radiation will show maximum cooling rates near the base of this stable layer. This effect would seemingly cause reduction of stability below (and enhancement of stability within) the stable layer. However, such a radiational influence would be felt as strongly during the day as at night. The small absorptivity of solar radiation by water vapor would tend to reduce the cooling rate slightly but this effect alone would appear to have a small influence upon lapse rates in the layer under consideration.

At any rate, clouds cannot be left out of the picture for they are typical of the trade wind regime (mean and modal low cloud cover is about 3 tenths at Eniwetok). Again, quantitative values are elusive, but the direction of influence of clouds upon the radiational heat exchange of their environment is well known. A cloud whose depth approaches 50 m or more can be considered a black body for terrestrial radiation. The cloud top thus radiates much more energy than it receives from the dry air above the stable layer and this upper 50 m or so of cloud cools steadily. In fact the deeper its penetration into the stable layer the less downward radiation it receives since moisture typically drops off rapidly with height.

Each individual cloud has a lifetime approaching a half hour. The upper radiational cooling is thus not likely to cause much change in the development of that particular cloud. But, the effect may be cumulative since the entire cloud layer is destabilized progressively by the effect of each cloud within it. This might allow each successive cloud to develop to a slightly greater depth. Under favorable conditions a few

hundred feet of cloud depth gained in this way might well spell the difference between success or failure of the coalescence mechanism to run its course.

It remains of course to demonstrate that this chain of events is suppressed in daytime. Sufficient absorption of solar energy by the cloud tops would contribute to this both by tending to evaporate cloud droplets, thereby retarding coalescence growth, and by reducing the instability of the cloud layer. Quantitative measures of absorptivity of clouds for solar radiation may be insufficient to allow a definitive computation of this effect.

It must be kept in mind that the high reflectivities of cumulus clouds are the result of volume scattering. This results in a very large intensity of radiant energy in the upper portions of clouds and a rapid decrease of this intensity with depth. Thus with a constant absorptivity (fraction of energy absorbed to that incident upon a slab) most of the absorption is accomplished by the upper layers of the cloud.

From the measurement of total absorption of cloud layers of various depths (Fritz and MacDonald, 1951) it would be reasonable to expect a cumulus cloud of 2 km depth to absorb at least 10% of the energy incident upon its top. Let us assume that 3% is absorbed by the upper 100 m of cloud. This is not inconsistent with the theoretical results of Hewson (1943). Assuming all of this energy is used to raise the temperature of the cloud mass, the rate of change of the temperature, T, with time is given by:

$$M_c \cdot \frac{dT}{dt} = \frac{dQ}{dt} = .03 I_o \dots\dots\dots (2)$$

where  $M$  = mass of upper 100 m of cloud

$c_p$  = specific heat at constant pressure of cloudy air

$I_0$  = rate of incident radiation per  $\text{cm}^2$

$Q$  = heat added

Since the clouds will be continuously forming and dissipating, the effect of this absorbed energy will be to provide heat to the layer of clear as well as cloudy air at that elevation. Since typically about 20% of the layer may include cloud at one time the heat must be distributed over 5 times the area over which it is received. We have then

$$M = 5 \times 10^4 \text{ cm}^3 \times 10^{-3} \text{ gr cm}^{-3}$$

$$c_p = .24 \text{ cal gr}^{-1} \text{ } ^\circ\text{K}^{-1}$$

and 
$$\frac{dT}{dt} = \frac{.03 \times 1.7}{50 \times .24} \text{ } ^\circ\text{K min}^{-1} = .26 \text{ } ^\circ\text{K hr}^{-1}$$

Even if allowance is made for eddy diffusion of this heat upward and downward, a heat source of this magnitude must be considered significant. This is true since the lower portions of the cloud layer are benefitting very little from such heating. This differential heating can easily change the lapse rate within the layer by over  $\frac{1}{2} \text{ } ^\circ\text{C Km}^{-1}$  in 6 hours. Under the delicate balance which exists in the oceanic convection layer the result upon individual cloud buildups could be substantial. The effect would be cumulative, reaching a maximum after noon, and would allow long-wave radiation to provide cumulative destabilization of this layer throughout the night.

It must be stressed that there is no reason a priori to expect that the variation of low cloud amount would parallel that of shower frequency,

although some interaction could be expected. Without increasing the number or size of clouds substantially, a large increase in rainfall probability could be achieved. The cloud depth is the most critical factor. Eventually, increased shower activity might well result in more scud-type cloud fragments and more persistent stratiform "aprons". The cloud variation might then be expected to lag the shower variation if the latter were the primary effect. Riehl (1947) in seeking an explanation for the early morning maximum in low cloud at Atlantic ship stations, attempted to test a radiation hypothesis. He sought differences in the nocturnal cloud increase between nights with considerable middle or high cloud and nights with little or none. His results could not lend support to any radiation mechanism. No attempt was made to repeat the experiment with Eniwetok data. In the development given here, however, it is not necessary to involve the variation of low cloud amount in the mechanism for rain production.

In meteorology it generally seems to be a mistake to look myopically upon a single factor to explain a complex phenomenon. Other influences can be found which may contribute in the same sense as the mechanism described here. The most important of these is probably the diurnal variation of sensible heat transfer across the air-ocean interface. Even with the "rootless" convective regime described earlier as being representative of this region, any change in the rate of heat transfer at the lower boundary must carry some influence to the cloud layer. However, it is more likely that a slight increase in ocean surface heating would result primarily in more cloudiness and have only a secondary effect on showers.

Diurnal changes in sensible heat transfer at the ocean surface are

admittedly small, but under tropical conditions these may be important. The main determinant of this transfer is, of course, the air-sea temperature difference. Extended periods of accurate monitoring of this variable have been rare, in the tropics particularly. Interesting results from a 16 day, "on station" cruise by the Woods Hole vessel, CRAWFORD, at about  $11^{\circ}\text{N}$ ,  $52.5^{\circ}\text{W}$  have been reported by Garstang (1958). Detailed measurements of air and sea temperatures revealed that on the average, maximum upward sensible heat transfer took place between 0500 and 0600 hrs. From 0900 to 1700 hrs the transfer was actually negative, that is from air to sea. This state of affairs resulted primarily from the observation that the air diurnal temperature variation as measured aboard ship was large ( $2.7^{\circ}\text{F}$ ) compared to that of the sea surface ( $1.1^{\circ}\text{F}$ ). Thus even though the sea is slightly warmer than the overlying air in the mean, the air becomes warmer than the sea toward mid-day. A similar effect may be found in the data collected by the METEOR Expedition (Kuhlbrodt and Reger, 1936). Radiation errors in air temperature measurements during daylight could contribute to this result. However, Jacobs and Clarke (1943) observed the same effect aboard the CARNEGIE even after correcting for radiation error, although the sea-air temperature difference usually remained slightly positive with a minimum before noon. The explanation for the greater diurnal range in the air temperature probably rests upon the active participation of surface air in the radiational transfers. It cools by night more rapidly than the sea surface (with its large heat reservoir) and likewise absorbs sufficient solar radiation by day to increase its temperature at a rate significantly greater than that of the sea.

In summarizing these conjectures dealing with the diurnal variations

of cloudiness and precipitation over the sea, it seems that the absence of a nocturnal maximum would be surprising. Only under light wind conditions could convincing arguments be made for an afternoon maximum of cloudiness and possibly of precipitation also. It has been suggested here that the primary mechanism leading to an early morning maximum in shower activity is the cumulative effect of nocturnal cooling of ever-present cloud tops. This leads to a steady destabilization of the cloud layer which causes individual clouds to have maximum development toward daybreak. The effect may be self-limiting because of increased convective mixing. Therefore the maximum activity may occur well before sunrise. Also, during the night upward transfer of sensible heat is increasing from the lower boundary with the result that clouds are more easily formed and somewhat more numerous near sunrise. This latter effect probably has little direct influence on the precipitation potential of each cloud, however. Shortly after sunrise both of these trends are reversed. The upward flux of sensible heat decreases rapidly, reducing slightly the occurrence of cumulus cloud elements (after some lag). The solar radiant energy absorbed by the cloud tops is slowly spread through much of the upper portion of the cloud layer, and the layer becomes steadily more stable. The reduction in cloud depth in turn limits the precipitation mechanism.

#### Temperature and humidity

Radiational heating and cooling of thermometers have some influence upon diurnal temperature variations measured within a standard instrument shelter. This is also true to a lesser extent for relative humidity and wet-bulb temperature. The extent of the error so introduced will, of course, depend upon the quality of the shelter, its exposure and the technique used

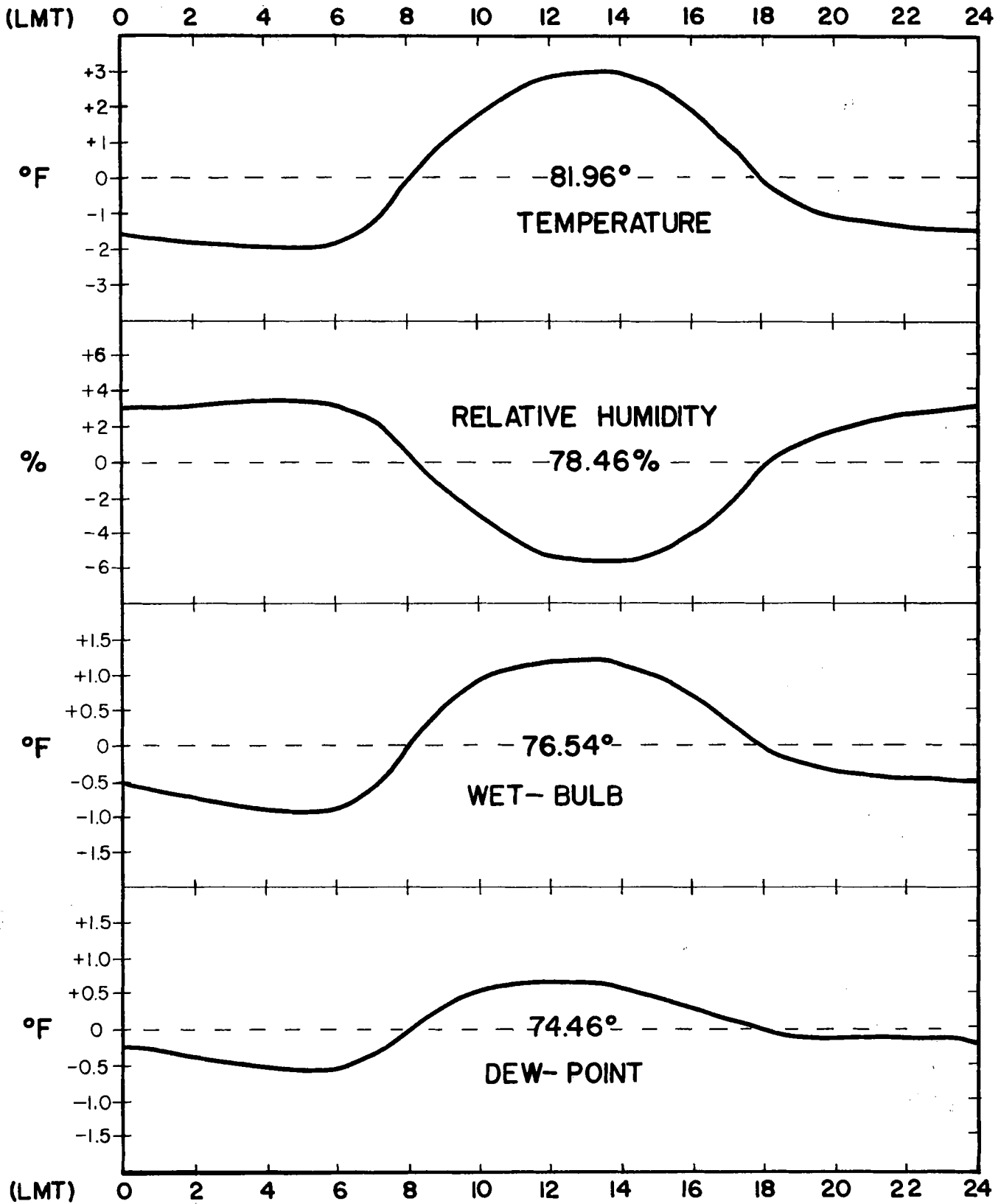


Fig 14. Annual mean diurnal variation of surface temperature, relative humidity, wet-bulb temperature, and dew-point at Eniwetok. Curves connect hourly mean values expressed as percent deviations from annual means (values shown).

in taking readings. None of these factors is known for the Eniwetok data. It is likely that the standard white, wooden, louvered thermometer screen was used and that it was located over coral fill. If a good fan driven ventilation system was used in the shelter or if psychrometer readings were taken in ventilated shade, it is likely that the true daily temperature range is not exaggerated by more than 1 or  $1\frac{1}{2}$ °F. The only humidity measure which should not be affected by such errors is the dew-point temperature ( $T_d$ ).

The annual mean diurnal variation of temperature and of three humidity variables is portrayed in Fig 14. The first four harmonic components of each of these curves will be found in Table 2 (page 13).

The diurnal temperature range is seen to be about 5°F. This can perhaps be compared with the range of 2.7°F reported by Garstang (1958) as appropriate to shipboard observations at the same latitude in the Atlantic. Results of Cruise VII of the CARNEGIE (Jacobs and Clarke, 1943) at these latitudes, largely in the Pacific, show a range of 3.1°F. Radiation errors in these values are probably of the same magnitude as those at Eniwetok. The influence of the atoll, then, upon the air temperature at  $4\frac{1}{2}$  ft measured at Eniwetok Islet is probably such as to cause warming of about 1.5°F at mid-day and perhaps less than 0.5°F cooling during early morning. These rough estimates are supported by a short analysis (not included here) made of the raw data presented by Blumenstock and Rex (1960) for a short study at Eniwetok.

The diurnal variation of wet-bulb temperature ( $T_w$ ) and of relative humidity (R.H.) conform to expectation, at least as far as phase is concerned. Table 2 reveals exact correspondence between the phases of the

diurnal components ( $t_1$ ) of  $T$  and  $T_w$  while that of R.H. is just reversed. The semi-diurnal components ( $t_2$ ) of  $T_w$  and  $T$  reach maxima within 25 minutes of each other; at the same time the 12-hour component of R.H. reaches its minimum. The foregoing would be the case even if the vapor content of the air did not vary diurnally.  $T_d$ , however, should measure this latter variation faithfully (once the pressure change effect is acknowledged to be negligible). The dew-point curve demonstrates that the moisture content of the lower air follows the temperature curve very closely. The diurnal range of about  $1.2^\circ\text{F}$  for  $T_d$  corresponds to a range of about 1.2 mbs for vapor pressure or approximately  $0.8 \text{ gr Kg}^{-1}$  for mixing ratio. The range in dew-point computed from mean data on  $T$  and R.H. provided by Garstang (1958) from ship measurements is about  $0.5^\circ\text{F}$ . The mean diurnal range of vapor pressure in this region of the Pacific shown in the records of the Carnegie Cruise (Jacob and Clarks, 1943) is about 0.7 mbs. We might then conclude that about half of the indicated diurnal range of moisture content at Eniwetok is due to the atoll influence. Presumably the soil and reef waters are sufficiently warmed to cause increased evaporation near mid-day compared to the open ocean.

It may be somewhat surprising to find so little correlation between the curves in Fig 14 and the diurnal rain frequency curve (Fig 10). There is no doubt that temperature and humidity respond very strongly to the presence of a shower; apparently the period of record is sufficiently long and showers are infrequent enough to allow the curves in Fig 14 to be essentially independent of rain occurrence.

Rawinsonde information

The Eniwetok card file utilized in this study included rawinsonde data at standard pressure levels for the 6 year period, January 1954 through December 1959. This card deck was processed for diurnal variations which might elaborate or clarify some of the problems which have been discussed for surface measurements. For this limited objective, only low levels were examined, and the results presented here are not designed to make a contribution to the interesting problem of diurnal variations in the free atmosphere.

Soundings at Eniwetok are normally made at 12 hour intervals. During three separate periods of nuclear testing, however, soundings were routinely made at 6 hour intervals with frequent special observations. These test periods each lasted from 2 to 6 months. About halfway through the period of record the standard hours of soundings were changed from 0300 and 1500 GMT to 0000 and 1200 GMT. The card file as a whole, then, covers the hours of the day fairly well in spite of its inhomogeneity.

In order to extract some idea of diurnal trends from this record the cards were examined sequentially, and forward differences were taken of the variables of interest. Only 6 or 12 hour intervals were accepted for differencing, however. When the interval between soundings was 12 hours (+ 1 hr) the change in the variable was divided by two and this value was assigned to the median hour. Thus, the change in the value of a quantity which is recorded at 0800 hrs could be due to differences between 0200 and 1400 LMT soundings or to the observed change from a pair of soundings taken at 0500 and 1100 LMT. In either case the recorded difference would be appropriate to a change per 6 hours.

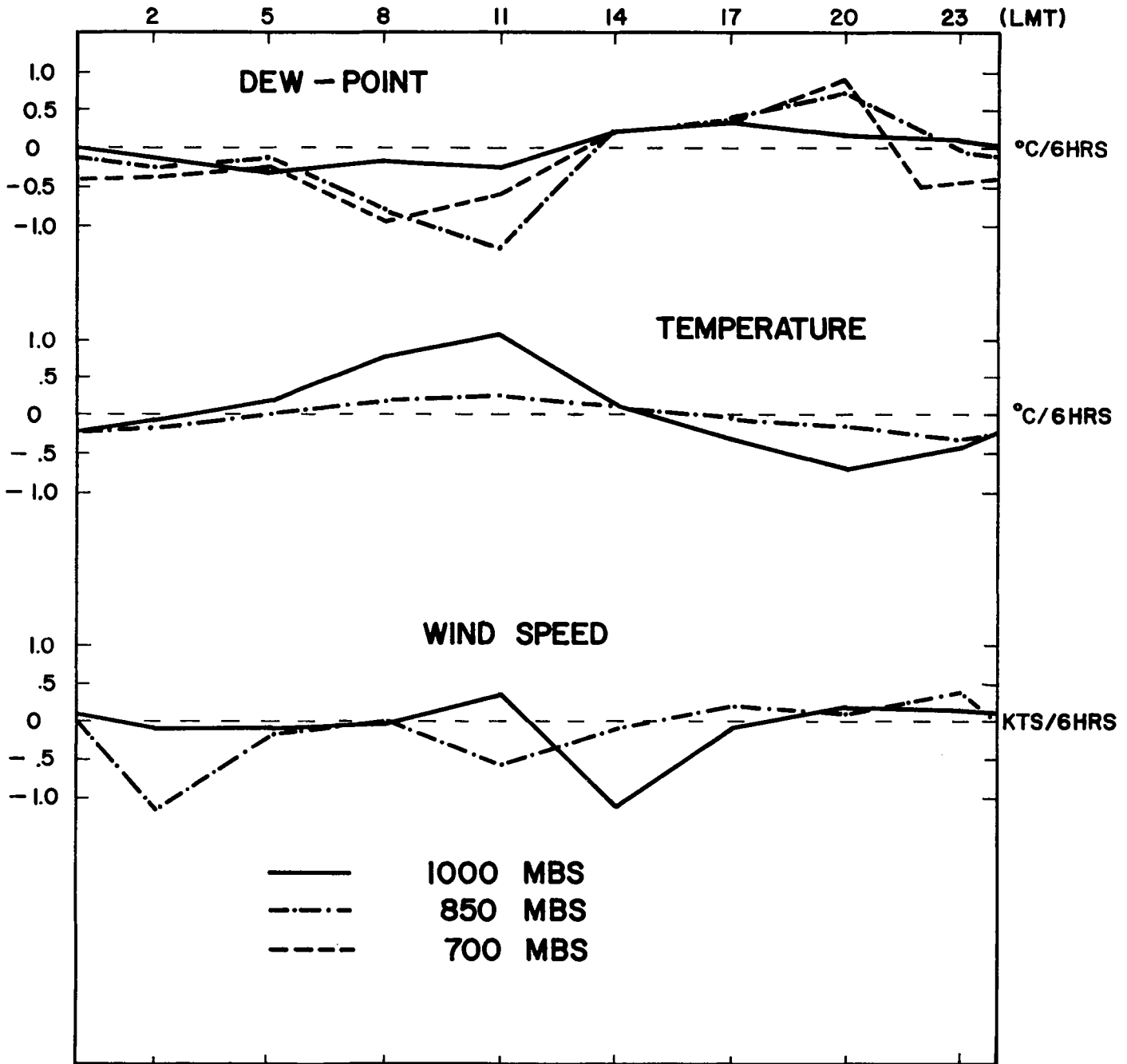


Fig 15. Annual mean diurnal variation of the rate of change of temperature, dew-point, and wind speed (per 6 hrs) in the lower atmosphere over Eniwetok.

The results of this analysis are presented in Fig 15 in graphical form for temperature, dew-point and wind speed. The dew-point variation was analyzed at 1000, 850, and 700 mbs while temperature and wind speed were examined at only the two lower levels. Since the curves depict the trend of the rate of change of the variables, the times of maxima and minima are to be found at the points where the curves cross the zero axis toward negative and toward positive values respectively. For convenience these times have been interpolated from the graphs and listed in Table 5. Also included in the table is an estimate of the range of the diurnal variation at each level, arrived at by integrating the positive and negative portions of each curve separately. Values have been entered for the surface layer (Fig 14) for reference.

TABLE 5

Variation with Height of Diurnal Trends of T and T<sub>d</sub>

	Temperature			Dew-Point Temperature			
	<u>Sfc</u>	<u>1000mbs</u>	<u>850mbs</u>	<u>Sfc</u>	<u>1000mbs</u>	<u>850mbs</u>	<u>700mbs</u>
Time of max.	1330	1500	1630	1200	2400	2300	2200
Time of min.	0500	0300	0500	0500	1300	1400	1300
Range (°C)	2.8	1.3	0.5	0.7	0.6	1.2	1.1

The vertical variation of the diurnal temperature curves follows expectation (see e.g. Harris, 1955; Harris et al, 1962). The time of maximum temperature is known to lag with height, and the range should decrease with height. Furthermore the curves demonstrate little radiational influence since radiation errors would tend to increase with height and to induce a temperature maximum at noon. It must be admitted, however, that the observation of a smaller temperature range at 850 mbs (about

5000 ft MSL or 3200 ft above cloud base) than at 1000 mbs does not appear to be in harmony with the hypothesis advanced earlier to explain the nocturnal shower maximum. It was there suggested that the primary nocturnal destabilization occurred in the cloud layer rather than the sub-cloud layer. If this were true it would be expected to result in a relative maximum in diurnal temperature range near the cloud tops (approx. 6300 ft) where radiational effects are largest. This would seem to be necessary if the diurnal change in lapse rate were to have the correct sense at these levels. The question, of course, cannot easily be resolved and requires considerably more information for satisfactory analysis. Data from intermediate levels and a consideration of other aspects of the heat budget (particularly convection and water vapor aspects) are essential.

The diurnal variation of the dew-point temperature (or mixing ratio) is similar at each of the three standard levels; the maximum value is found near midnight and the minimum in early afternoon. The author is unaware of any systematic instrumental errors at work here, but there are two reasons to suspect that this result is unrealistic. Firstly, the  $180^{\circ}$  shift in phase from the surface to 1000 mbs (approx. 300 ft) cannot be easily accounted for. Secondly, one would expect a noticeable forward tilt of the humidity wave with height through the lower 10,000 ft of the atmosphere in accordance with studies over land. Still, our ignorance of the details of convection processes over the sea and their diurnal variation preclude a positive conclusion. It is conceivable, for example, that the surface maximum results largely from decreased upward transport during mid-day when the lowest air is most stable. The occurrence of a minimum at upper levels would follow through similar reasoning. The reverse distribution would be expected during the night. This reasoning, however,

would lead one to expect the maximum at high levels to be reached at the time of greatest convective activity, i.e. just before sunrise rather than near midnight. Further conjecturing is unwarranted here. A strong need for micro-meteorological measurements over the ocean is certainly evidenced.

Whereas individual monthly mean diurnal trends for temperature and dew-point conform very closely to the annual means portrayed in Fig 15, the wind speeds were highly variable. Little confidence can be placed in the lower curves of Fig 15. For this reason the information on wind speed was not included in Table 5. The analysis nevertheless indicates the absence of strong, organized diurnal variations aloft. The hours showing greatest speed change with time, 0200 and 1400 hrs, are just those hours where the accuracy is poorest (least number of observations). It will be noted, nevertheless, that the 1000 mb curve shows a semi-diurnal variation with maxima at 0100 and 1200 hrs. This is in fair agreement with the surface data. The 850 mb winds show a maximum at 2330 and minimum at 1530 with a range of about 1 kt.

Part III: Synoptic-Scale Disturbances.

Considerable progress has been made in recent years in the elucidation of the nature of perturbations of the trades. The early model of the "easterly wave" disturbance has generally withstood the pressure of synoptic case analysis and theoretical scrutiny. With more detailed observations, however, exceptions to the basic model were certainly to be expected and these have appeared. The question of geographical or seasonal limitations for these types of disturbances has not been explored. Meanwhile, the role of upper-level circulations in the trade regime has been stressed repeatedly, most recently in Pacific case studies made by the Woods Hole Oceanographic Institution (Riehl et al, 1959; Malkus et al, 1961). The subtropical cyclone which lowers its influence into the trade wind regime from upper-levels has been described by Ramage (1962). Tiros satellite observations have showed organized perturbations in the regions (both at high and low levels) to be much more plentiful than previously suspected (Sadler, 1962).

An elementary attempt is made here to investigate the climatological aspects of perturbations in the trade wind regime as exemplified at Eniwetok. It was hoped by so doing to bring out the nature of those disturbances which are most significant in terms of weather production in this portion of the Pacific. For this purpose a machine analysis was made of the  $9\frac{1}{2}$  years of hourly surface observations for two types of perturbations: those which disturb the field of wind direction, and those which disturb only the speed field.

Disturbances in surface wind direction

The punched card format at our disposal utilizes a sixteen point compass for specification of surface wind direction. As a first step in the analysis a simple classification of clouds and precipitation versus wind direction was arranged. This is a useful approach here because of the constancy of wind direction; winds from directions other than E and ENE are short-lived and almost invariably represent synoptic-scale disturbances of some kind.

All available observations (over 75,000) were classified according to wind direction, and tabulations were made of 7 quantities: 1) occurrence of precipitation in each of three categories: light showers, steady rain, and moderate to heavy showers, 2) the amount of low cloud cover, 3) the number of 6-hourly rain reports and the amount of rain recorded (6-hourly observations only).

A seasonal breakdown in such a tabulation is obviously desirable. In terms of the surface wind, the seasons at Eniwetok are fairly clear-cut. From November through June the prevailing direction is ENE, and speeds of less than 5 mph are very infrequent. From July through October the prevailing wind is East and the speeds are noticeably lower; the modal speed in each month is less than 15 mph, and light winds and calms are common. It will be noted that the above seasons correspond closely to the non-typhoon and typhoon seasons in the western Pacific. The 4 month "weak trade" season also corresponds to the period during which the "mid-Pacific trough" becomes firmly established at upper-levels over this region.

The tabulations are summarized in graphical form in Fig 16. The number of occasions with winds from the western semi-circle were so few as to give unreliable percentages, so that these (along with calms) are included under the column heading "other".

Some explanation of the method of representation in Fig 16 is in order. The top three graphs for each season represent the relative frequency of occurrence of each precipitation category under different directions of the surface wind. The "percent occurrence" refers to the proportion of all observations under that particular wind direction which report the particular weather category. Thus, when the wind is from SSE during the winter season there is an 8.7% chance that an airways observation will report a light shower occurring. At the same time the chances of steady rain and of moderate to heavy showers are 10.7% and 5.0% respectively. Thus the probability of rain in any form with SSE winds is 0.24, that is, the percentages are additive.

The graph describing the variation of low cloud amount is straightforward. The graph labeled "rain intensity" refers to the mean amount of rainfall reported under the indicated wind direction whenever measurable rainfall was reported. This graph then portrays rain amounts for "rainy situations" only and for this reason is labeled as intensity. The relative proportion of rainy situations, on the other hand, is given in the top three graphs. The last of the six graphs of each set is a standard histogram presentation showing relative frequency of each category of wind direction.

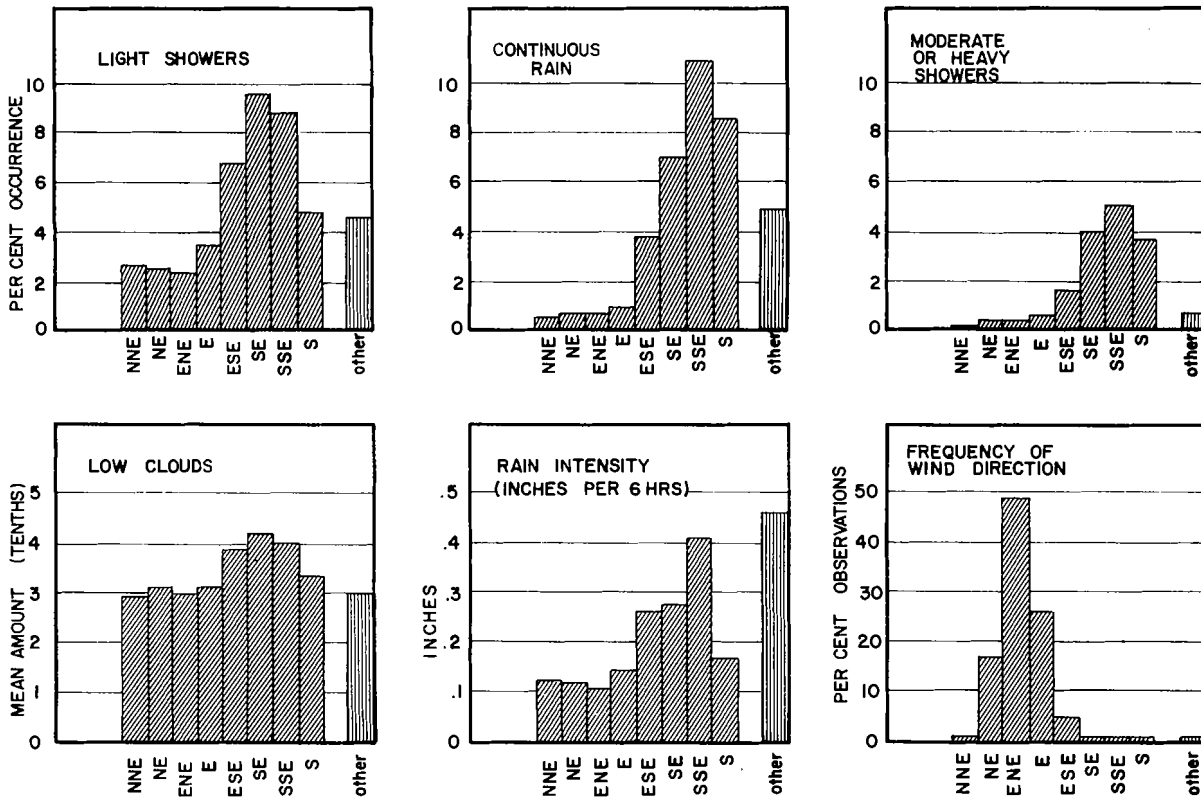
Certainly the most striking feature of these graphs is the marked

deterioration of weather associated with a surface wind south of East. This effect is more pronounced during the strong trade season. In either season the least chance of precipitation is to be found under ENE flow, but deviations toward southerly winds offer the greatest increase. During the winter, rain is almost 8 times more likely with SSE winds than with the prevailing ENE winds; moderate to heavy showers are about 15 times more likely. These figures are even more arresting when it is realized that some of the southerly winds arose during merely disorganized, light-variable flow.

At the same time it will be noted that southerly excursions of the wind are associated with greater low cloud cover, although the change is small. It seems obvious that the more frequent and heavier precipitation experiences with winds from the SE quadrant are the result of more unstable clouds rather than increased cloud cover. There would probably exist a good correlation between cloud depth and surface wind direction.

Deductions as to the nature of disturbances affecting Eniwetok made on the basis of this climatological evidence suffer from severe limitations. Nevertheless, winds from directions other than E or ENE occur less than 30% of the time and therefore must be largely representative of disturbances. We may conclude that in such disturbances the worst weather occurs under southerly excursions of the wind. Thus the results do not run counter to the standard easterly wave model but should perhaps not be particularized beyond saying that in most disturbances affecting Eniwetok equatorward flow yields near average or slightly better than average weather; poleward flow gives much worse weather. The above points can be made even more convincingly through Fig 20 which depicts

NOVEMBER-JUNE



JULY-OCTOBER

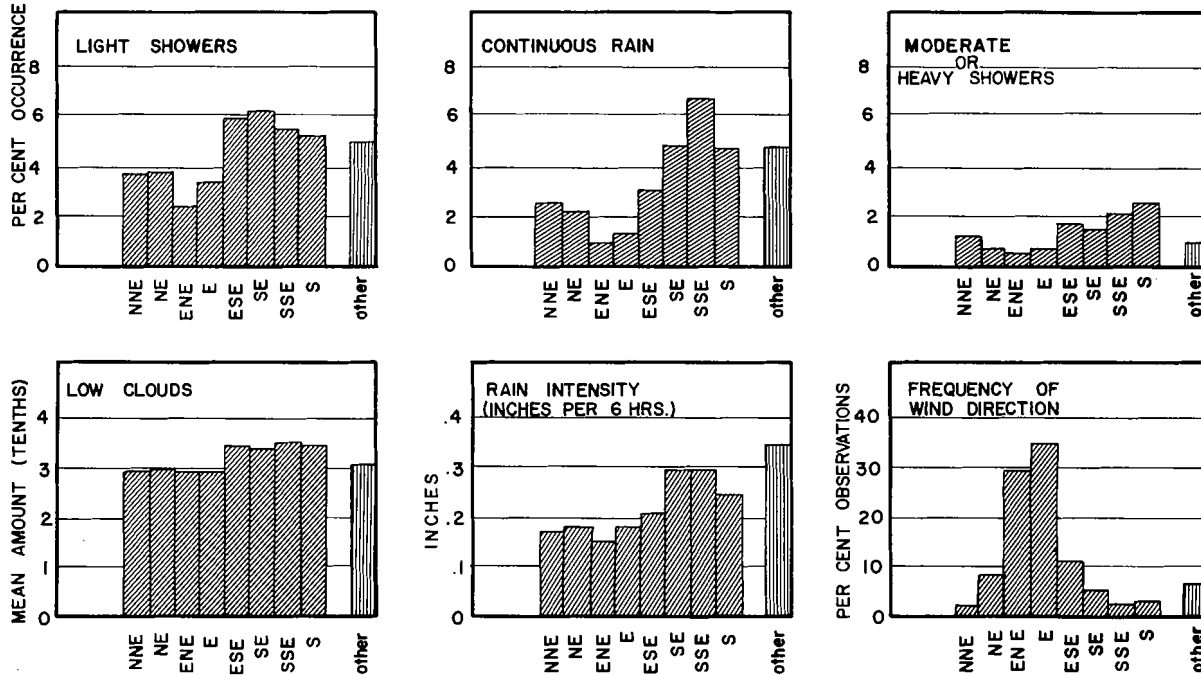


Fig 16. Weather as a function of the direction of the surface wind at Eniwetok within the strong trade wind season and within the season of weak trades. November-June comprise about 50,000 observations, July-October about 25,000.

the distribution of average rain amount with varying wind direction. The curves show the mean rainfall in each direction category expressed as a percentage of the overall mean. Prevailing winds show near normal rain amounts while more northerly winds show somewhat drier conditions. The seasonal difference is exemplified in the two curves.

Whether the disturbances affecting Eniwetok are of some "easterly wave type" or in fact result from the passage of closed vortices to the south can of course not be answered by these data alone. It is also possible that upper-level systems, weakly reflected at the surface, affect these results. Similar climatological studies conducted for other stations in the Marshall Islands and vicinity could very well offer considerable illumination. Lacking this, it may be profitable to examine the correlation of upper-level conditions with surface wind direction for hints of the vertical structure of the disturbances.

#### Soundings versus surface wind direction

All rawinsonde data cards from Eniwetok from January 1954 to December 1959 (6770 observations) were processed and classified according to the direction of the wind at the 1000 mb level. If the wind speed was less than 5 kts the wind was classified as "calm" and its direction was not used. A tabulation by 10 degree classes of wind direction was then made of dew-point values at 1000, 850, 700, and 600 mbs and of the heights of the 1000, 850, and 700 mb surfaces. The tabulations of these 7 variables were then grouped according to the two seasons described in the previous section. The mean values were computed and expressed as deviations from the overall seasonal mean (for all wind directions). The results are

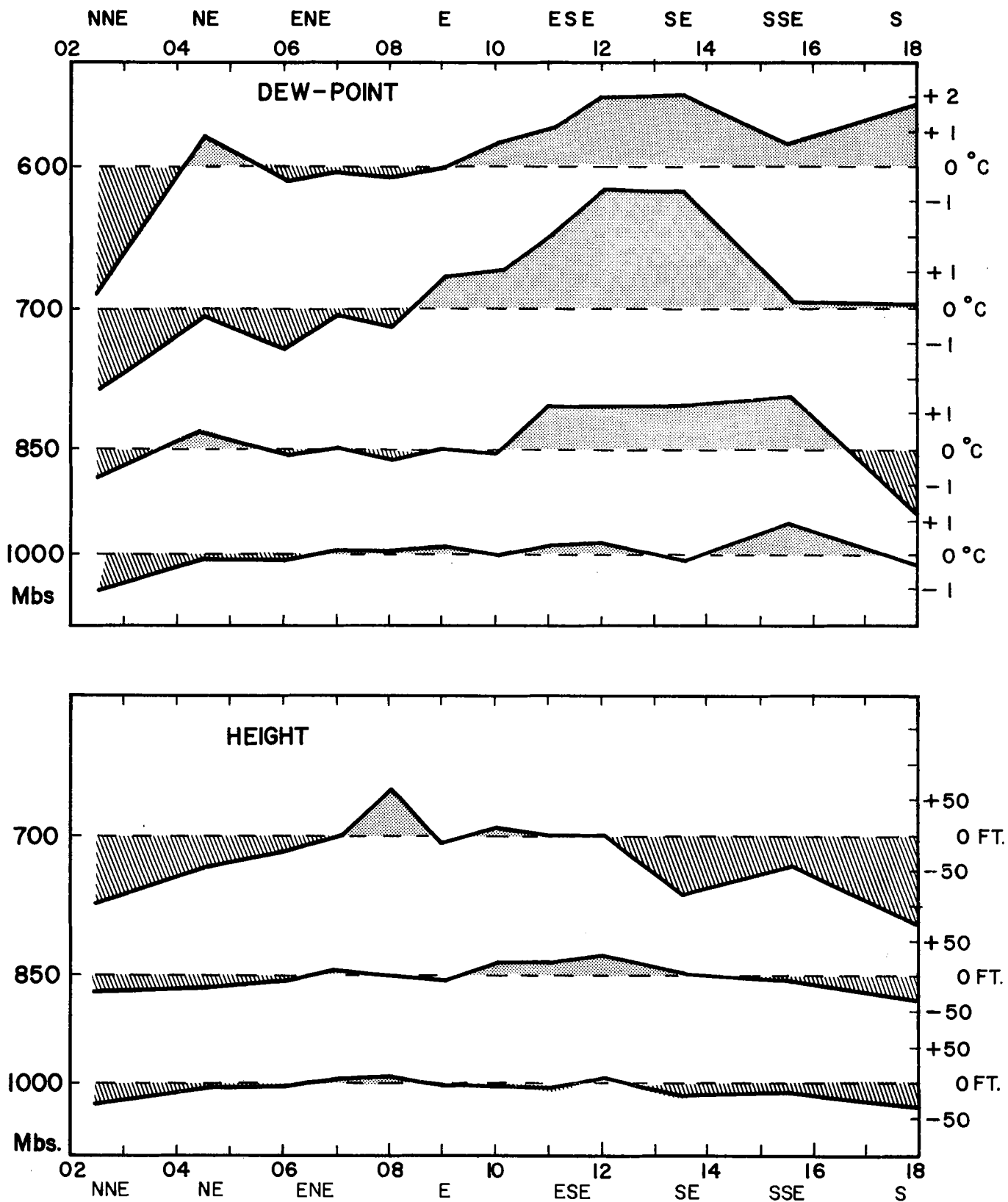


Fig 17. Dew-point and pressure-height in the lower atmosphere over Eniwetok as a function of the direction of the 1000 mb wind for the season of strong trade winds, November-June. Variation at each level is expressed as a mean deviation from the monthly mean at that level.

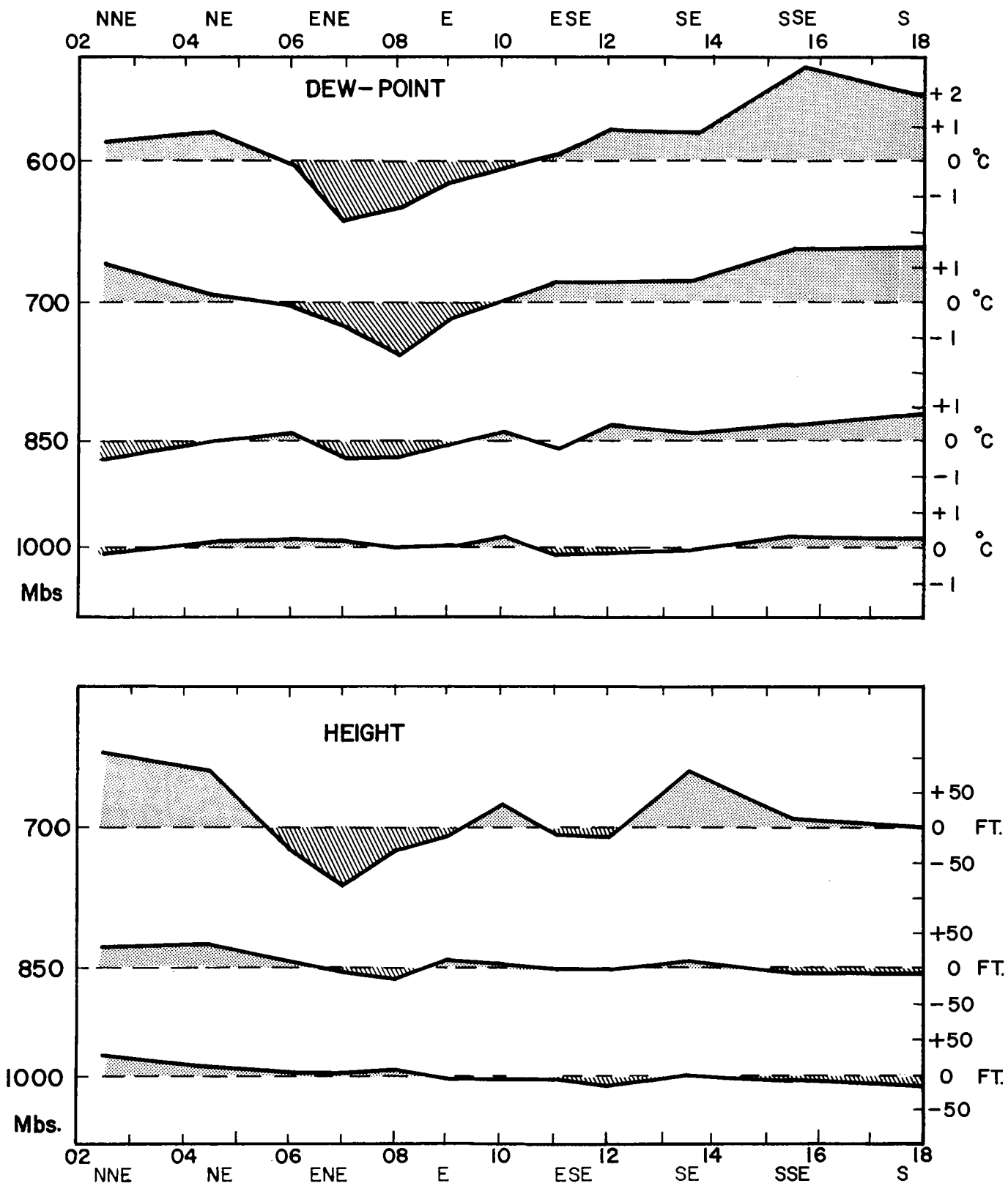


Fig 18. Same as Fig 17 but referring to the weak trade wind season, July-October.

displayed graphically in Fig 17 and 18.

Mean deviations have been entered on these figures at intervals of 10 degrees for wind directions between 60 and 120 degrees, but fewer observations of other directions made averaging to the points of the 16-point compass appear desirable.

The results of this analysis generally confirm the conclusions drawn in the previous section. The southerly excursions of the wind are accompanied by an increase in moisture aloft and therefore by a lifting of the stable layer limiting low cloud development. It should be noted, however, that the mean dew-point increase is relatively small, reaching a maximum of about  $3.5^{\circ}\text{C}$  at 700 mbs under SE winds during the winter season. This is equivalent to an increase of roughly 10-15% in mean relative humidity at these levels if the temperature change is negligible. Thus, in spite of the very large increase in raininess under these conditions the increase in upper-level moisture is only nominal. Still, the association is unmistakable. Fig 18 further demonstrates that the intensity of these perturbations is smaller during the summer season.

The variations of the height of the pressure surfaces depicted in the lower curves of Fig 17 demonstrate that excursions of the wind away from the prevailing direction are indeed associated with the passage of low pressure systems. The intensity of the "disturbances" appear to increase with height to at least 700 mbs. The height curves in Fig 18 representing the summer situation, on the other hand, fail to provide a convincing correlation with wind direction.

The correlation of upper-level moisture with wind direction is

displayed in Fig 19 in a different form. The percentage of radiosonde ascents which report "motorboating" of the humidity element (relative humidity less than about 12%) is entered for each of the wind directions previously discussed (southerly winds were too scarce in winter to allow a representative percentage computation). The variation is large and convincing only in winter, whereas the mean rain amount versus wind direction (Fig 20) shows a good relationship during both seasons. The seasonal difference in depth of the moist layer is well pointed up by Fig 19. This, of course, has great implications in the problem of typhoon formation in the western Pacific.

In summary, the analysis of dew-point and heights reported by rawinsondes at Eniwetok shows a good correlation with low-level wind direction, particularly during the 8 month "winter season". The data tend to support a conclusion that equatorward flow in disturbances is frequently associated with subsidence and poleward flow with general low-level convergence. There is some evidence that the disturbances are generally most intense at about 700 mbs.

#### Disturbances in surface wind speed

The resident within the trade wind regime is generally sensitive to slight changes in his environment. Variations in the strength of the trades are quite common and easily perceived. One naturally tries to correlate these changes with cloud and shower activity, and forecasters in these regions frequently have "rules of thumb" built up from such experience. On large or mountainous islands the intensity of sea-breeze effect and of orographic influences will obviously be affected by

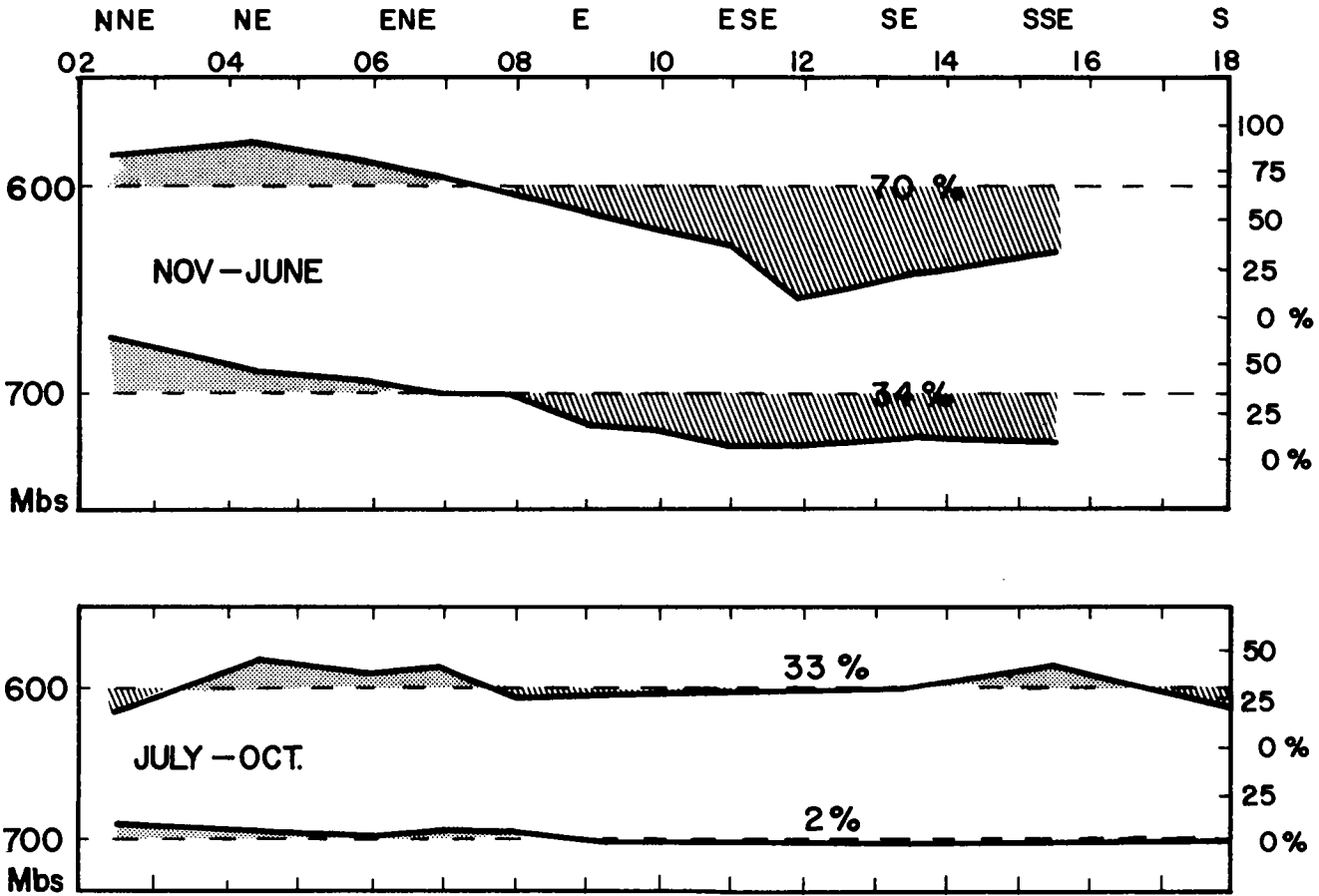


Fig 19. Percent of radiosonde ascents at Eniwetok which report humidity too low to measure (i.e. "motor-boating") at 700 and 600 mb shown as a function of wind direction at 1000 mb. Dashed lines depict the mean percentage for all ascents regardless of wind direction.

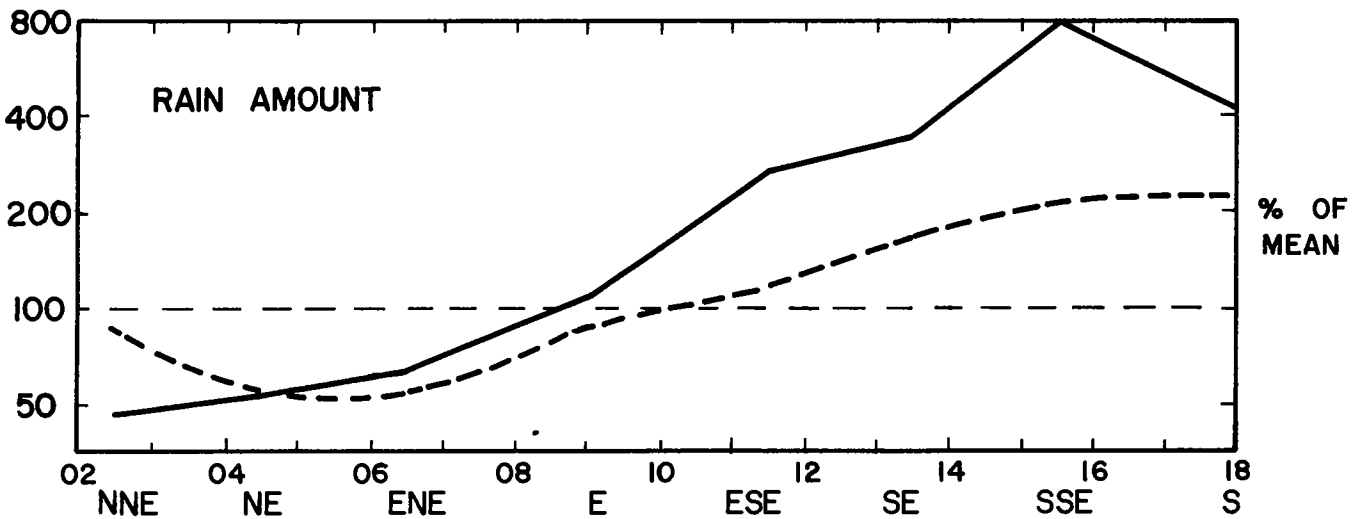


Fig 20. Mean rain amount reported at Eniwetok under varying direction of the surface wind. The means for individual directions (to 16 points) are expressed as a percentage of the grand mean regardless of wind direction. Solid curve refers to the strong trade season (November-June), dashed to the season of weak trades (July-October).

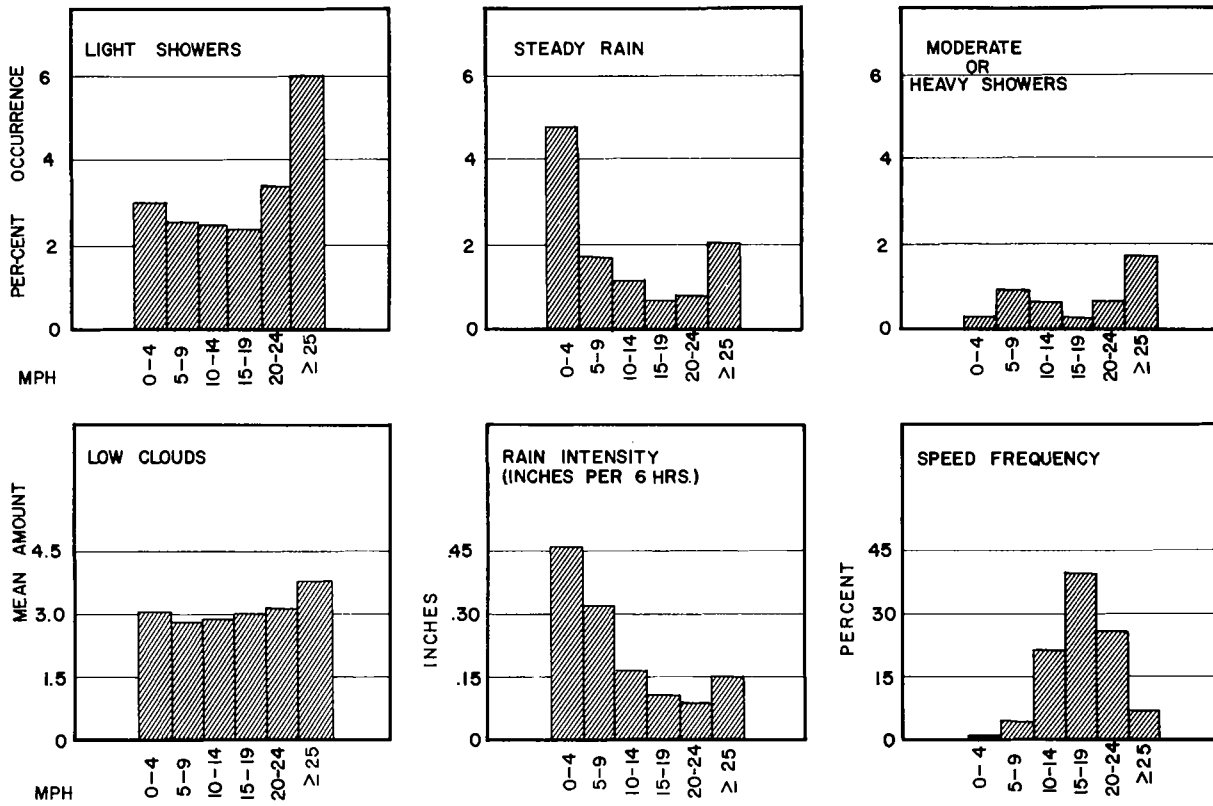
changes in the speed of the prevailing wind. But again the first question to be answered should deal with the nature of these disturbances (if they qualify for that term) over the open oceans, before local effects are introduced.

In order to develop a framework for such a study, Eniwetok surface data were examined for dependence of cloud and rainfall upon the speed of the surface wind, regardless of direction. The results are presented in Fig 21 in the same graphical format used for the wind direction analysis (Fig 16, p 59). The speeds are broken down into six categories of 5 mph width. The histograms showing the frequency distribution of speeds point up the difference upon which the seasonal distinction has been based.

The set of graphs demonstrate once again that the most common wind regime has generally fine weather while excursions away from the mode are associated with worse weather. An interpretation of these excursions as representative of disturbances is not made as easily here as with wind direction. Periods of unusually strong winds definitely seem to be associated with disturbed conditions in either season, but a seasonal distinction appears in the case of abnormally weak flow. Low winds speed do not give rise to more low cloudiness or rainfall than normal in summer, although they do appear as a definite concomitant to disturbances during the winter season. Individual months were tabulated in all cases to be sure that seasonal summaries were not merely statistical accidents. There was general consistency in all cases.

The seasonal differences in weather versus speed categories may be a reflection of real differences in the nature of disturbances in the two

NOVEMBER - JUNE



JULY - OCTOBER

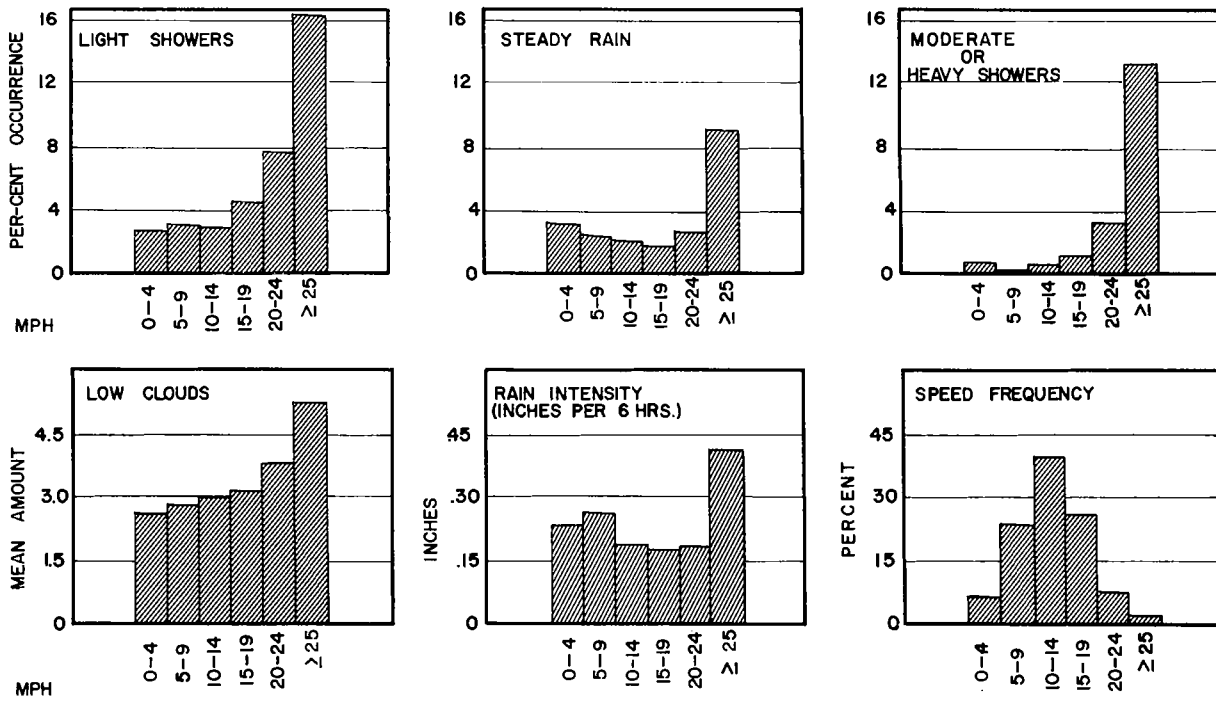


Fig 21. Same as Fig 16 but referring to the speed of the surface wind.

seasons, but conjectures are perhaps inappropriate at this point. It should be noted, however, that in winter, light winds are very unusual and therefore are more likely to represent a highly disturbed condition. Light winds are much more common during the summer and hence strong winds are more likely to be indicative of a synoptic disturbance.

Trade wind maxima and minima

The foregoing analysis made no effort to distinguish between perturbations in speed and direction of the surface wind. Indeed, the two are usually inseparable. Nevertheless, the easterly wave disturbance, for example, affects primarily the direction field in its early stages of development, and it would seem that a case may be made for synoptic scale disturbances whose effect on the trade winds may be largely in the speed distribution (initially at least). An isotach analysis in a typical trade flow is admittedly no easy task, but isolated maxima and minima are common features.

Variations in speed within a straight parallel flow much certainly set up patterns of divergence and convergence which could have important bearing upon weather distribution. The expression for the divergence of a field of motion in terms of distance "s" along a streamline and distance "n" normal to the streamline (positive to the left looking downstream) is

$$\text{Div} = \frac{\partial V}{\partial s} - V \frac{\partial \beta}{\partial n} \dots\dots\dots (3)$$

where V is the speed and  $\beta$  is the direction of the wind. If we may limit ourselves to flow which is essentially parallel, the second (diffluence) term becomes negligible. Then an isotach maximum with a speed gradient of 6 kts in 300 km would yield  $\text{Div} = 10^{-5} \text{sec}^{-1}$ . This is a respectable

magnitude in terms of vertical motion generation and weather production, provided that the value persists within the same air mass for sufficient time. However, directional diffluence amounting to about 25 deg per 300 km could yield the same magnitude of divergence so that care must be taken to rule out the well known tendency for compensation between these terms.

Lacking accurate knowledge of the spatial distribution of the wind speed, the most important parameter to be studied at a fixed point is obviously the rate of change of wind speed with time in cases of negligible directional change. Provided that these isotach centers move nearly in the direction of the wind we should expect increasing winds to be associated with convergence and decreasing speeds with divergence.

The surface data deck for Eniwetok was used to make a survey of the serial change in 6-hourly mean wind speed. It was hoped that such changes would be representative of synoptic scale effects. The 6-hour periods were chosen so that the last hour coincided with the regular synoptic observation during which rain amount is recorded, i.e. 0500, 1100, 1700 and 2300 LMT. Whenever two consecutive 6-hour periods each showed prevailing E or ENE flow (i.e. at least 4 of the 6 observations in each period were from these directions) the mean speeds were computed and compared. It was felt that such a definition of "undisturbed trades" would allow for the occasional minor fluctuations in direction (many times induced by showers) which appear in the data presumably as a result of the short averaging period used in reporting an airways observation.

The difference between the mean speeds in the consecutive 6-hour periods was used as a classification variable against which were

tabulated: 1) mean low cloud amount for the second 6-hour period;  
2) serial change in mean low cloud cover from period 1 to period 2;  
3) number of hourly rain observations in period 2; 4) rain amount reported for period 2. This tabulation was carried out for individual months of the year and averaged over 1 kt intervals of the serial change in mean wind speed. In making this tabulation it was recognized that heavy showers could themselves influence the wind speed in a given observation and give rise to a correlation which is extraneous to the subject of "synoptic scale" disturbances at which this study was aimed. In order to avoid any local influences of squalls, the magnitude of the wind was not included in the 6-hourly mean computation whenever moderate or heavy rain was reported in present weather.

The tabulation was printed out on a monthly basis, but close examination revealed no annual trend or basis for seasonal distinction. The results were combined and are portrayed in Fig 22. Curve A depicts the variation in mean low cloud amount during the second consecutive 6-hour period associated with the speed change shown as abscissa. A positive speed change signifies an increase in surface wind with time. The cloud values are expressed as a percentage of normal low cloud cover (2.93 tenths). It is seen that a good correlation exists between cloud amount and the magnitude of the speed increase. Large decreases in speed, however, also appear to be associated with somewhat more low cloud cover than normal. The variation in cloud amount is admittedly small (little more than one tenth of sky), but as a "measure of disturbance" it is fully as large as those discussed earlier in connection with southerly excursions of the wind direction.

### SERIAL CHANGE IN 6-HRLY. MEAN WIND SPEED

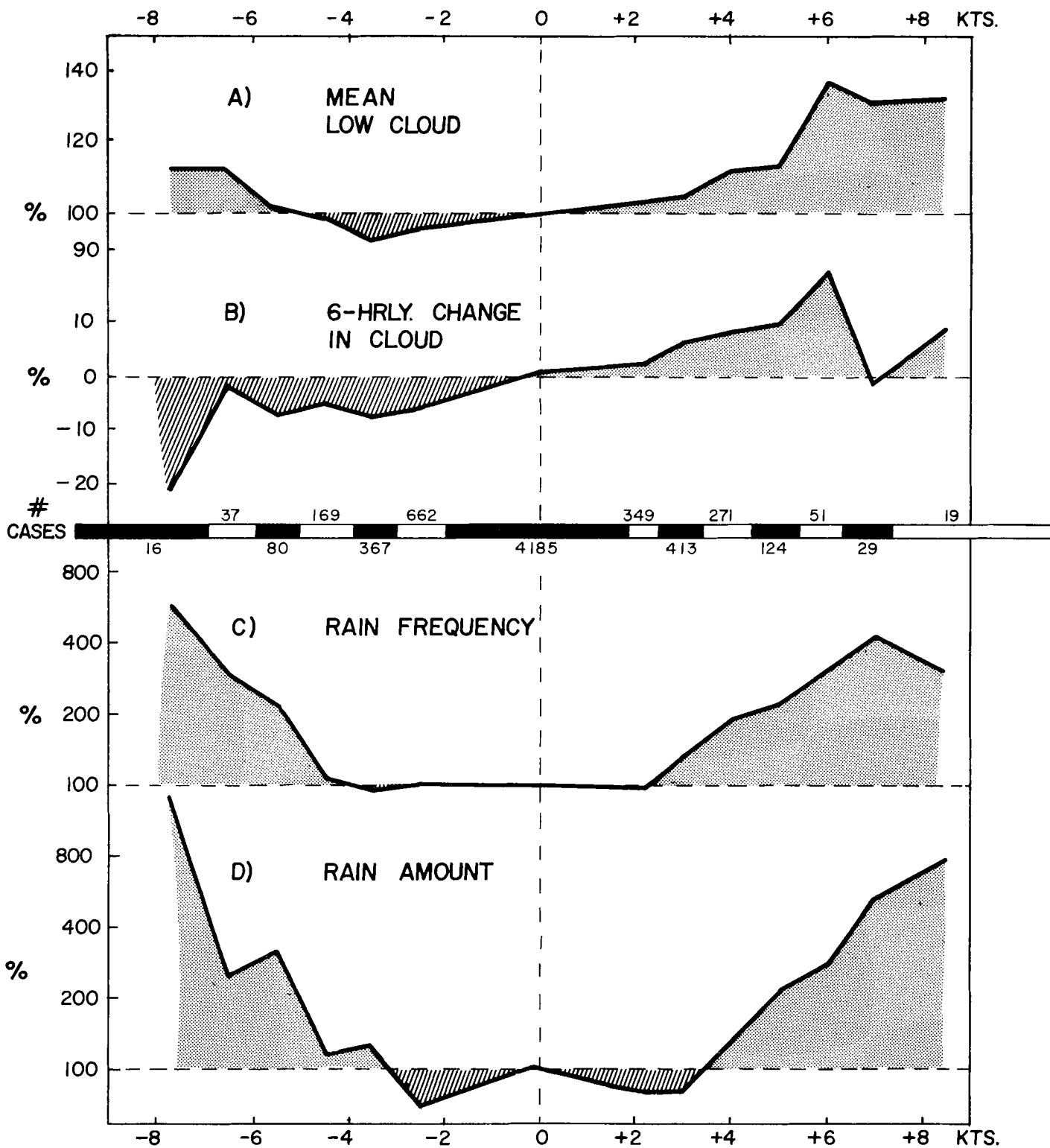


Fig 22. Weather as a function of the serial change in 6-hourly mean wind speed. Curves show mean variations expressed as a percentage of overall mean. Center scale shows averaging interval and the number of observations in each class.

In terms of relating these speed changes to the passage of fields of divergence and convergence, the serial change in low cloud cover from one 6-hour period to the other is of more interest. This is shown in Fig 22, curve B, where the changes are expressed again in terms of percent of normal cover of 2.93 tenths. A good correlation is evidenced indicating perhaps that speed increases are associated with low-level convergence and speed decreases with low-level divergence. However, care must be taken in drawing this conclusion, although it is what one would expect if the sign of the 1<sup>st</sup> term on the right side of Eq (3) truly controls the sign of the divergence. It would be more generally correct to conclude from Fig 22B that increasing wind speed with time is associated with increasing convergence and that decreasing speed usually indicates decreasing convergence (or increasing divergence).

Curve C shows how the mean number of rain observations reported during the second 6-hour period varies with the magnitude of the "speed disturbance". The values are again expressed as a percentage of normal. Surprisingly, higher rain frequency occurs with both increasing and decreasing speeds. The same sort of picture is seen in curve D which portrays the variation of mean rain amount reported at the end of the second 6-hour period. The mean rainfall for these cases of extreme increase in speed was some 8 times normal; with the cases of extreme slackening of the wind, rainfall was 15 times normal.

In terms of the premise advanced at the beginning of this section, the bad weather associated with slackening wind might appear anomalous. There are several possible explanations for this result, including development of upper-level disturbances, isotach centers which move

upstream, or situations in which the confluence term in Eq (3) is not negligible and actually overbalances the stretching term. Of these three alternatives the last is most reasonable. It seems likely that sharp decreases of wind speed with time are more usually associated with isotach maxima moving away from the station than with approaching speed minima. Under this condition one might expect the region of greatest convergence to be found near the isotach maximum where both components of Eq (3) would probably contribute in the same sense. On the upstream side of the isotach maximum the confluence term would predominate but the convergence would decrease with distance upstream. In other words, the speed (stretching) term tends to predominate ahead of an isotach maxima and the direction (confluence) term is most important behind these features. It is just in such cases as these that the confluence is least likely to show up in the analysis of the wind at a single station. It is quite possible then that the wet periods indicated at the left end of curves C and D of Fig 22 result from situations in which confluence predominated over the stretching effect in spite of the attempt to minimize the former.

The question of the relative frequency of these speed disturbances can be partly answered by referring to the scale in the center of Fig 22. The alternate heavy and open zones demark the extent of the speed scale over which values were averaged to form the curves. The numbers refer to the absolute frequency of these events in an 8 year period at Eniwetok. The extreme cases appear to be very rare (>6 kts in 6 hrs occurs 9 times per year), but it should be stressed that this analysis made no attempt to pick out all such disturbances. No overlapping periods were examined;

a change which largely straddled the static 6-hour periods or extended over more than one period would be suppressed.

It can certainly not be claimed that this brief study has demonstrated the existence of disturbances manifested primarily in the speed field of the lower trades. However, the evidence must be called at least suggestive and would seem to warrant a more concentrated attack on the problem. It is recognized that disturbances such as these must develop curvature within the flow pattern eventually (should they persist), but there appears to be no valid reason to suspect that speed disturbances could not be an important weather phenomenon at low latitudes. A closely related problem in applied tropical meteorology involves the degree to which the sign of the convergence is controlled by confluence in the streamline analysis. Many meteorologists equate the two terms for practical purposes. Further climatological analysis along the lines presented here, but using a network of stations with spatial as well as time derivatives, could throw light on this important problem.

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GLOSSARY OF TERMS

Airways observation: coded weather observation reported regularly on the hour by airport weather stations. It is meant to depict weather variables affecting aircraft operations; valid for 3 to 5 minute period ending on the hour.

coalescence mechanism: the process by which a few raindrops grow from a large population of tiny cloud droplets, involving growth by collision. Once a few larger droplets are produced, their greater fall speed enables them to sweep out more and more cloud droplets and thereby grow at an ever-increasing rate.

colloidal stability (of a cloud): a measure of the tendency of cloud droplets to remain in suspension, i.e., not to aggregate into precipitation-sized droplets.

conditional instability: a distribution of temperature with height which allows slight vertical displacements to be stable only on the condition that the air be unsaturated. If "conditionally unstable" air becomes saturated, a parcel of air displaced upward would become buoyant and accelerate upward as a result of the release of latent heat by condensation.

confluence: the opposite of diffluence; see diffluence.

convergence: the net accumulation of mass of air with time within an arbitrary (small) area on a horizontal map. This air must then be moving either upward or downward at this point since the mass per unit volume (density) remains essentially constant. If the level in question is near sea level, the air must be moving upward. Thus, low-level convergence is associated with rising air and generally bad weather.

diffluence: the condition characterized by the tendency of streamlines to move apart. In this case the air tends to flow away from some axis. This does not necessarily lead to a depletion of mass in the region (divergence) since the variation of speed of the flow must also be considered. See divergence equation.

divergence: opposite of convergence. Near the earth's surface it is associated with sinking motion of the air and relatively clear skies.

$\text{Div} = \frac{\partial V}{\partial s} - V \frac{\partial \beta}{\partial n}$  : a partial differential equation describing the magnitude of the divergence in terms of the quantities on a horizontal streamline-isotach map. The symbol,  $\partial$ , indicates an increment in the value of the following variable. Thus the first term on the right describes the contribution of the rate of change of speed per unit distance along a streamline (stretching effect). The second term is a product of the speed

at a point on a streamline and the rate at which the streamlines change direction per unit distance along a line normal to the streamline at that point. This latter rate of change describes the tendency of the streamlines to diverge and is referred to as the diffluence effect. Both of these effects contribute toward divergence, the opposite tendency would lead to convergence.

harmonic analysis: a procedure for expressing the periodic variation of some quantity in terms of the set of oscillations (cosine functions) which best represents the variation. The first cosine curve (first harmonic) covers the fundamental period (here, 24 hours) with one maximum and one minimum. The second cosine curve (second harmonic) has two maxima and two minima in the same period and so forth until the sum of the derived cosine curves exactly duplicates the observed variation. The amplitudes of the "best fit" harmonics generally differ, and their magnitude indicates their relative importance in describing the daily trend of the quantity.

inversion: an atmospheric layer in which the temperature increases with altitude in contrast to the normal "lapse" with height. The "trade wind inversion" is common to the subtropics, particularly in the eastern oceans; its altitude varies from about 2 to 8 thousand feet. Its importance lies in the fact that it forms an effective upper limit to low cloud development. In the trades, then, the air is typically very dry above the inversion. Downstream this inversion becomes higher and less distinct but upper-level dryness tends to persist.

isopleth: line on a graph or chart connecting points of equal value of some quantity which varies over the chart.

isotach: line connecting points of equal wind speed, usually referring to horizontal wind speed on a weather map.

lapse rate: rate at which temperature decreases with height in the atmosphere.

mixing ratio: mass of water vapor per unit mass of dry air, usually expressed in terms of grams of vapor per kilogram of dry air.

motorboating: name given to the condition under which the humidity element in the rawinsonde fails because of very low relative humidity. The term describes the telemetered signal received through a monitoring speaker in such a situation.

rawinsonde: radio transmitting device hung from a buoyant balloon which telemeters to a ground station a nearly continuous record of temperature, humidity, and pressure as it ascends. The instrument is tracked by radio direction-finding equipment or radar in order that horizontal winds can be computed from the balloon drift. If used without wind measuring equipment, the instrument is called a radiosonde.

sensible heat: heat which can be transferred by conduction as a result of a temperature difference (also called enthalpy).

standard pressure level: the reference points in a vertical sounding of the atmosphere at which, by international agreement, the rawinsonde information is always recorded, coded and transmitted. These reference levels are identified by the pressure reading of the rawinsonde.

streamline: line which is parallel to the wind direction at every point along its course. A set of properly spaced streamlines on a horizontal chart thus portrays the instantaneous flow pattern.

subsidence: gentle descending motion, generally through a deep layer of the atmosphere and over a large area.

synoptic: usually applied to the analysis of large numbers of simultaneous weather observations. Hence, synoptic features are those depicted by weather maps --- weather systems on a scale of the order of 500 km or more.

thermal: more or less distinct mass of less dense (warmer) air which rises under buoyancy forces -- may or may not reach condensation level and continue rising as a visible "cloud thermal".