

from the  
space science and  
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the university  
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madison

ON THE REQUIREMENTS OF A BOUNDARY LAYER INSTRUMENTATION SYSTEM  
FOR THE GARP TROPICAL EXPERIMENT

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Final Report on the Boundary Layer Instrumentation Study,

Task Order Number 3 to STAG Contract E-127-69 (N)

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## SUMMARY

A boundary layer system capable of meeting the requirements of the GARP Atlantic Tropical Experiment must be able to--

Operate on a near continuous basis, under all but hurricane conditions, in a harsh and confined shipboard environment.

Take measurements which are representative of the entire boundary layer--up to 1500 m for the central and western Atlantic.

Return winds of sufficient accuracy to infer cluster scale divergence and convergence.

Define vertical structure in detail sufficient for computation of total boundary layer convergence and divergence.

Substantial variation on vertical and horizontal scales within the boundary layer necessitates time averaging on the order of tens of minutes, and measurements at a minimum of three levels in the vertical. During disturbed conditions, six to ten levels of observation may be needed for adequate vertical resolution.

Required observations and their accuracies are as follows:

wind speed $\pm 1\%$	pressure $\pm 0.1$ mb
wind direction $\pm 1^\circ$	humidity $\pm 2\%$
temperature $\pm 0.1^\circ$ C	

It is highly desirable that altitude be independently observed, with an accuracy of  $\pm 2$  m.

Systems based on double theodolites, rawinsondes, and rockets are incapable of providing at reasonable expense the near continuous, time-averaged, extremely accurate measurements which are required. Because tethered balloons can provide continuous observations over time periods approaching a day at any level up to the limit of their lift, the system best able to meet the observational requirements consists of multiple instrument packages attached to the tether of a helium airfoil balloon. Under normal operation these packages would be maintained at fixed levels. When conditions required greater vertical resolution the balloon and its packages would be lowered and raised in a stepwise cycle.

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I. Introduction

A. The Setting

The Global Atmospheric Research Program (GARP) is an international cooperative research effort whose objectives are --

- a. To obtain a set of global observations of atmospheric parameters of state good enough to serve as the initial conditions for definitive tests on atmospheric general circulation simulation models.
- b. To conduct a number of regional observation programs aimed at extending our understanding of the physics of the atmosphere in order to improve the performance of the models.
- c. To develop better mathematics - physics simulation or parameterization schemes for implicitly incorporating complex processes in the numerical models.

A major weakness of currently available numerical simulation models is their inability to properly handle tropical motions, particularly those associated with deep convection. Satellite observations have shown that convection is not random in the deep tropics, but is organized into zones of intense activity which appear in satellite pictures as cloud clusters. This convection occurs on a horizontal scale smaller than that which is used in typical numerical simulation models. The intense transport processes in cloud

clusters must, therefore, be parameterized so that the effects of sub-grid scale motions can be accounted for by the model which operates on grid scale synoptic observations.

If the cloud clusters are to be adequately parameterized, it is important to establish the co-spectra between vertical motions and the parameters of heat, moisture and momentum. Of key importance in any parameterization scheme is an adequate knowledge of the instability criteria which control the onset of cloud cluster formation and decay. Without these criteria, it will be very difficult to develop an effective cloud cluster parameterization.

At present, there are different views on how such instability arises. There are, of course, possibilities for upper and lower tropospheric motions which can give rise to convective instability. However, in the deep tropics where the coriolis parameter is small but its rate of change with latitude maximum, frictional convergence in the lowest layers of the atmosphere can occur in a zonal current with no horizontal shear. Suitable shear increases such convergence. This mechanism for stimulating the release of convective instability in the typical tropical atmosphere has been labeled boundary layer pumping, convective instability of the second kind, convergence of the sub-cloud layer, and so on.

A key component of the GARP Atlantic Tropical Experiment is a program to observe the boundary layer of the atmosphere well enough to determine instability criteria for convection, and to observe its general behavior in steady and disturbed conditions.

These observations must be obtained over the open ocean away from the influence of land. Even islands are not suitable platforms because of the possibilities of sea breeze effects. Since aircraft cannot satisfy the

requirement for extended observations, the observing platforms must be suitably equipped ships.

B. Scientific Requirements of Boundary Layer Observations for GATE

The Report of the Fifth Session of the Joint Organizing Committee of GARP, which summarizes the meeting held in Bombay from 1 to 5 February 1971, describes the requirements for observations of the boundary layer during GATE as follows:

... special emphasis must be placed on the sub-cloud layer, which will require additional instrumentation (e.g., tethered balloons). The sub-cloud layer observing system must have the capability to make observations in disturbed weather.

The primary objectives of these observations are:

- (i) To delineate the vertical fluxes of heat, vapor and momentum in the boundary layer within the B-scale area when the disturbances pass through the network concerned.
- (ii) To determine the typical pattern of these fluxes on the scale of the cloud clusters. This part of the study includes understanding of the role of the boundary layer in the convective instability of the second kind.

... the boundary layer measurements should be conducted at maximum operating intensity when the ships are within or near the 'cluster box'. It could also be of importance to know the transport properties of individual cumulus cells and even the convection of smaller scales (c and d scales).

The committee suggested that an observational programme from dedicated ships in the B-scale network incorporate regular fixed level measurements within lower atmospheric and upper oceanic layers. The measurements should...extend where possible to at least 1000 m...and to 1500 m if practicable.

Attached as an Annex to the JOC Report is a Report of the Second Session of the JOC Study Group on Tropical Disturbances, Geneva, 11-16 January 1971. This Study Group recommended that winds of the B-scale network be measured to one meter per second accuracy. However, "the budgetary computations necessary to determine the cluster truths require a greater accuracy than needed for large scale flow patterns and vorticity calculations."

In summary we state that measurements in the boundary layer must be accurate enough to establish the bulk transports on the B-scale of mass, moisture, heat, and momentum within the boundary layer in order to establish to what extent boundary layer pumping can influence the bulk properties of the troposphere above the boundary layer.

In the paragraphs which follow we summarize evidence on the space and time variability of the boundary layer. This variability will have an important impact on the design of the instrument system needed for boundary layer studies. If boundary layer behavior is very steady, only a few samples with time and with height will be required. On the other hand, if the behavior of the boundary layer is unsteady to an extreme, continuous samples at many heights will be required. This will have a large impact not only on the complexity and cost of the instruments and data system but on the operational cost as well.

## II. The Tropical Boundary Layer

### A. Space and Time Variability

The spacing of ships, which with the exception of aircraft and constant level, free floating balloons defines a minimum horizontal scale for sampling the boundary layer, is determined by typical dimensions of cloud clusters. Atmospheric fluctuations of smaller scale cannot be defined; to the extent that such fluctuations exist, they constitute a noise which for cluster scale studies should be averaged out.

Much of our limited knowledge about the space scale of small disturbances in the deep tropics comes from cloud studies. On the basis of ground and aircraft observations Kuettner (1959) stated, "Cumulus lines stretching in the general wind direction are a characteristic feature of convection over the

tropical oceans occurring specifically in the trades and in tropical cyclones." Malkus and Riehl (1964), in a series of airplane flights across the tropical North Pacific, found among tropical cloud distributions "a high degree of organization on many different scales." They were able to distinguish between longitudinal and transverse rows of cumuli, the former being by far the most common.

Satellites and manned space craft in the last decade have extended the findings of Kuettner and Malkus and Riehl. They show that lines and cells are most typical of the subtropical highs. Elsewhere there is great diversity of scale and pattern.

We might suppose that the band structure of tropical clouds would be reflected in space and time series measurements of wind, temperature, and moisture in the boundary layer. Two of the earliest attempts to measure the vertical and horizontal structure of the tropical boundary layer are described by Malkus (1957). The results are summarized as follows:

The trade wind moist layer is itself subdivided in the vertical into two superposed layers of different convective regime, because of the occurrence of water vapor condensation at about 650 m above the tropical oceans (Fig. 1). Below the condensation level, in the so-called "subcloud" layer, unsaturated convective turbulence predominates. Eddies 50 - 150 m across are characteristic and recent studies suggest that larger scales of motion with dimensions 10 - 50 km...are also significant. No evidence of cloud-scale motions below cloud base has been found, except in precipitating downdrafts. Above the condensation level cumulus convection is the major transport process; small-scale turbulence is confined to the neighborhood of clouds, which form in bunches separated by wider weakly subsiding clear areas.

The division of the tropical boundary layer into subcloud and cloud layers, together with variability in scale of cloud organization, may explain why time series of surface wind do not in general reflect organization of the atmosphere on small scales. Hwang (1970) presented a power spectrum of surface wind at



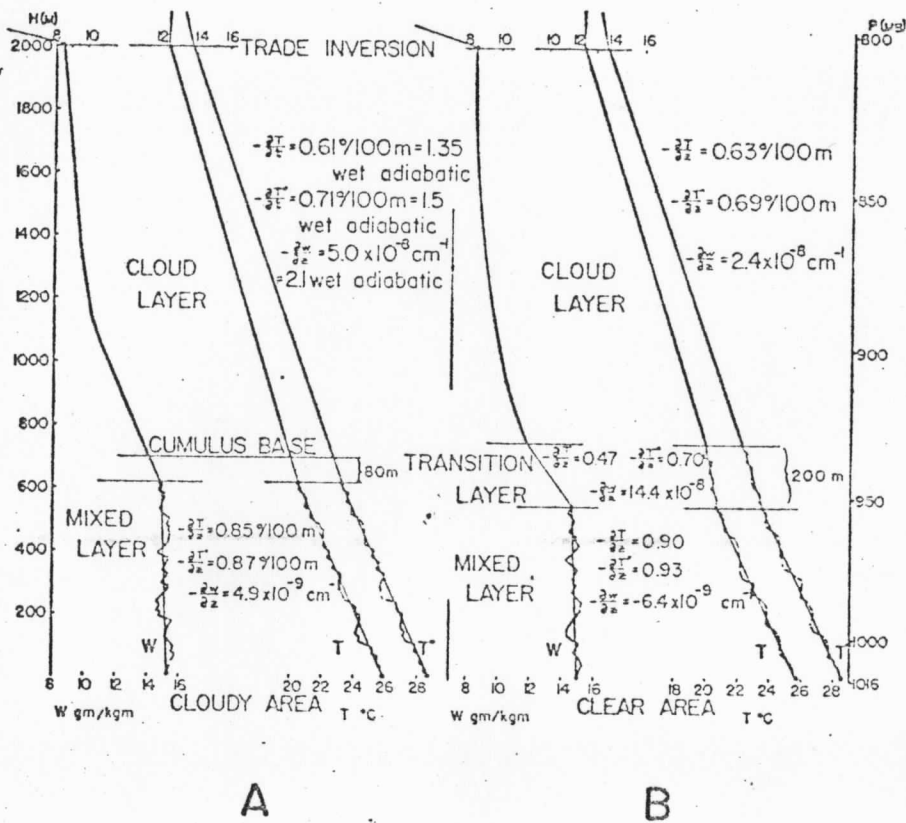


Fig. 1. Average soundings for the April 1946 Woods Hole expedition during a strong trade, compiled by taking the average temperature, T, virtual temperature, T\*, and mixing ratio, w, at the base of each layer, the average lapse rate within the layer of each property and the average vertical thickness of the layer. Height in meters is the ordinate. Fig. 1A is the average cloudy area sounding (compiled from nine individual soundings) and Fig. 1B is the average clear area sounding (compiled from sixteen individual soundings). Generally one or more soundings of each type were made on a given observing day (Malkus, 1957).

Palmyra Island (6°N, 162°W) in the central Pacific (Fig. 2). These data, collected during the 1967 Line Islands Experiment, showed no spectral peak from  $\frac{1}{2}$  to 100 cycles per hour under clear skies. In rain, as the ITCZ passed, there was a strong peak at about 7 minutes. Unfortunately, the spectrum at periods of less than an hour was based on only one hour of data for clear sky conditions and one-half hour of data for rainy conditions.

Fiedler and Panofsky (1970) commented on this absence of strong energy peaks as follows:

Over the sea, the situation is less clear. A spectrum by Millard (1968) shows an extremely strong gap (Fig. 3). However, the spectrum at high frequencies is suspect due to buoy motions. On the other hand, a spectrum by Frenzen (1970) based on unpublished data obtained during BOMEX suggests no mesoscale gap over the ocean. There are other reasons to suspect the absence of such a gap. On the one hand, microscale turbulence is generally weak over the ocean due to its small roughness. On the other hand, the larger variability of hourly means suggests the presence of structures, probably longitudinal rolls, which have periods on the order of an hour. It is likely that such structures are often broken up over land but may persist over smooth terrain.

The evidence therefore suggests that organization on space scales corresponding with periods of minutes to an hour will be present intermittently. Frequency of occurrence will depend on location with respect to the major synoptic features, weather conditions, and possibly altitude within the boundary layer.

Constancy is a trademark of winds in the tropics. Within the trade wind zone Riehl (1954), Ficker (1936) and other investigators have found that vertical profiles typically have a low level "nose" and little change of direction with height (Fig. 4). Kuettner (1959) quotes Riehl and Malkus (1957) on the reasons for such two-dimensional profiles.

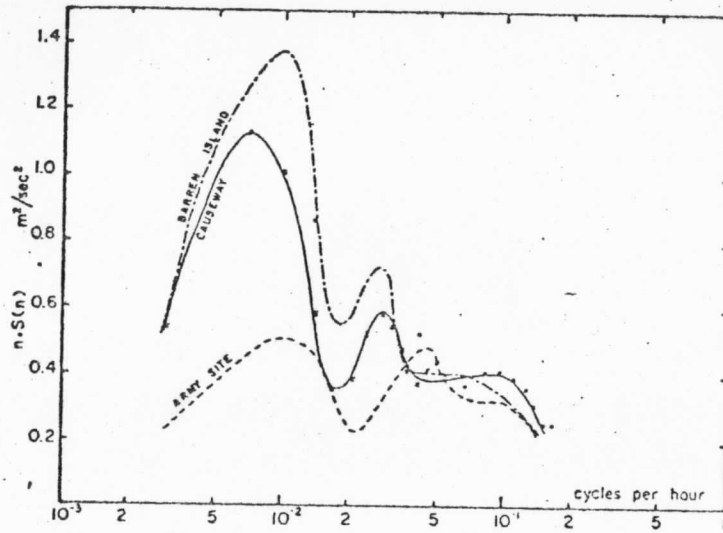


FIGURE 2a Power spectra of surface wind speed recorded by MRI anemometer. The solid line is for causeway data, the dot-dashed line for Barren Island data, and the dashed line for Army site data.

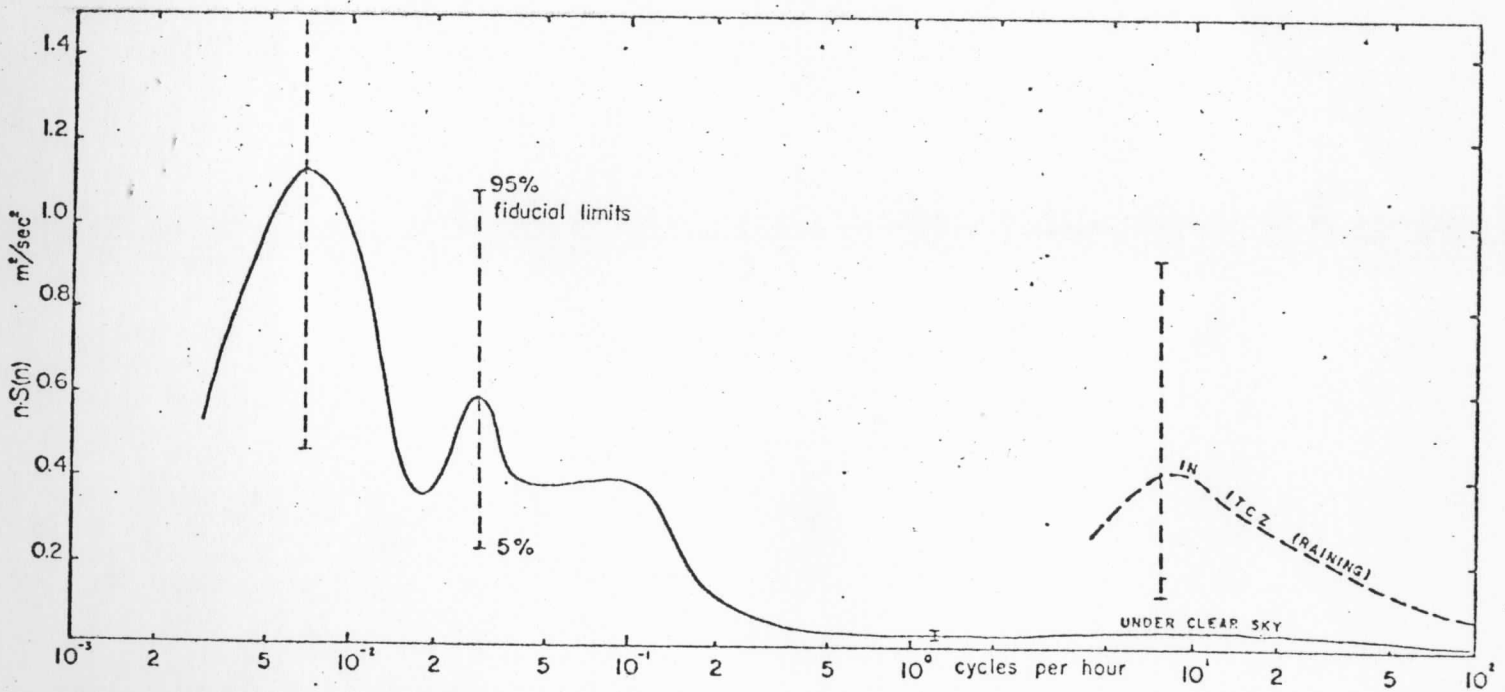


FIGURE 2b Power spectra of surface wind speed at the causeway, Palmyra Island. The dashed line was based on data from a short period of precipitation during an ITCZ passage. (Hwang, 1970)

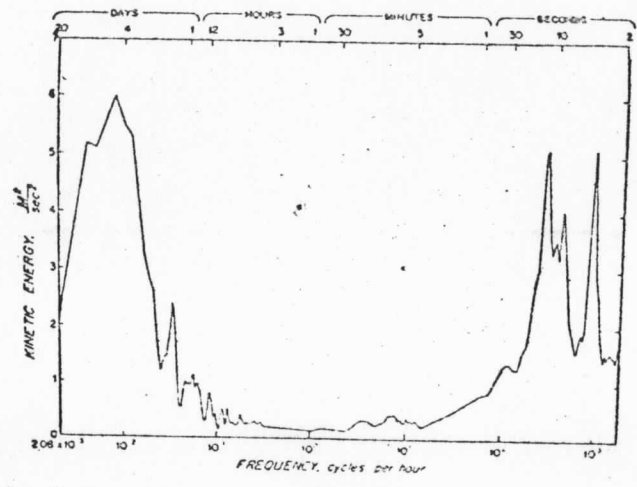


FIG. 3. Composite spectrum of wind speed over the sea for periods from 20 days to 2 sec (Millard, 1968).



From a viewpoint of balance of forces [Riehl and Malkus] postulate "a vertical wind profile with a curvature in the turbulent layer such that turbulence must act to retard the flow...this kind of profile is typical of the portions of the trade wind belts where the meridional temperature gradient is directed poleward. The geostrophic wind is strongest at the ground and decreases throughout the tradewind layer. On account of friction, however, the maximum wind is situated some distance above the ground. In the trade section (along a ground streamline) it is warmer at the downstream than at the upstream end. Since the pressure decreases from right to left across the section looking downstream, the isobars will rotate counterclockwise with height under the influence of the temperature field and thus assume more and more the direction of the surface wind.

However, Riehl and Malkus caution that because this structure depends on the "existence of a trade inversion, and a vertical wind profile with a curvature in the turbulent layer such that turbulence must act to retard the flow...the conclusions drawn cannot be applied to the tropical and equatorial regions in their entirety."

That variations do occur, even in areas considered to be within the trades, is evident from the daily mean profiles of Charnock, Francis, and Sheppard (1956) made at Anegada Island, 150 km east of Puerto Rico (Fig. 5). Substantial day to day as well as vertical variations of direction and speed were recorded. The average of these profiles (Fig. 6) shows a continued change of direction up to 1350 m, the limit of observations. About half of this change was attributed to the thermal wind.

Only in thermalwind corrected averages over large samples do winds typically exhibit an Ekman variation with height. Individual profiles, according to Gray (1968), are affected by gust scale turbulence (100-500 m), the effect of which is large relative to the frictional component. Observational errors further increase profile variability, but these effects can be eliminated by averaging if they are random, and Gray states, "There is no reason to think they are systematic."

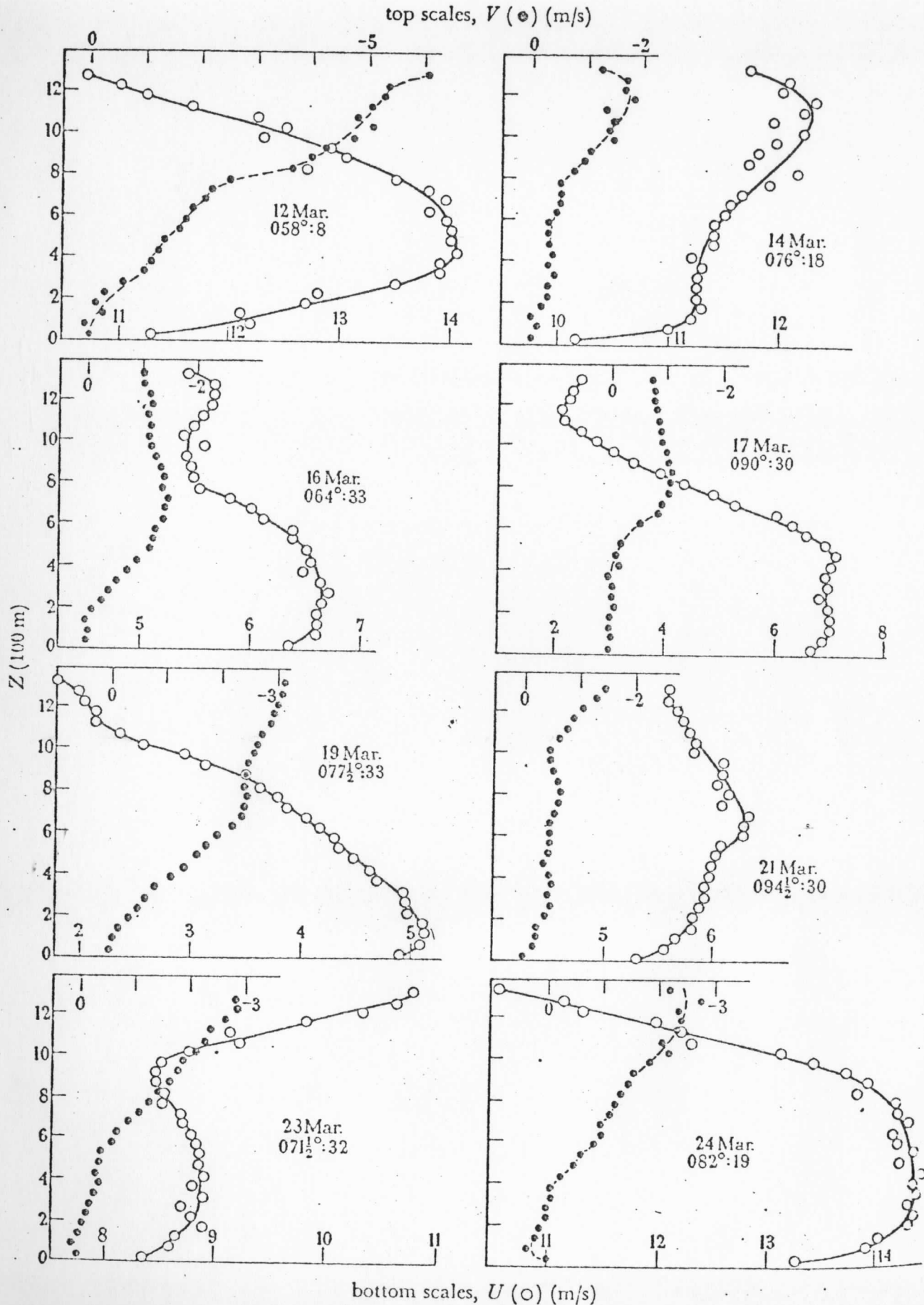
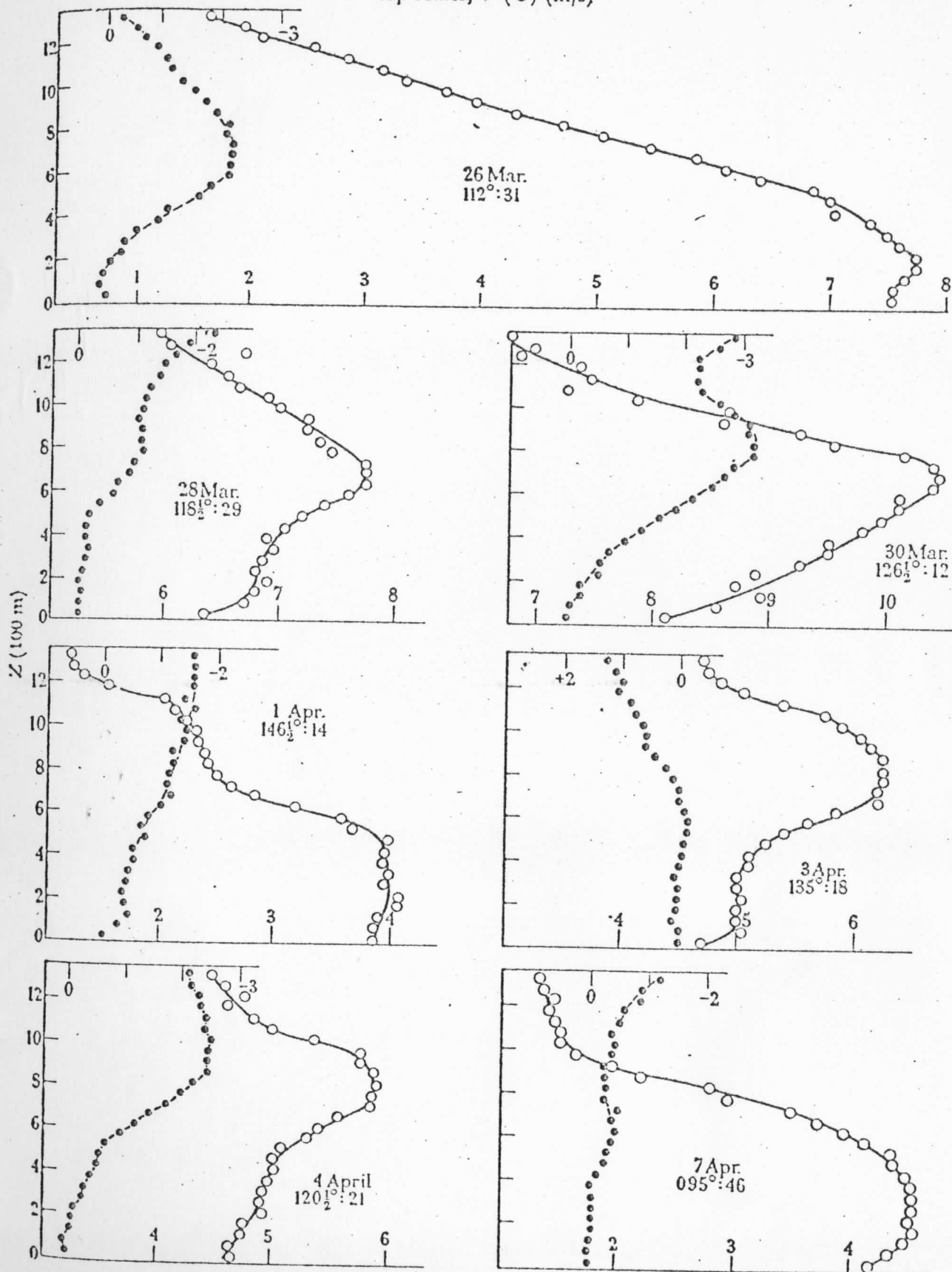


FIGURE 5. Daily mean profiles of wind components  $U$  ( $\circ$ ) and  $V$  ( $\bullet$ ). The direction of the surface wind in degrees from true north, and the number of ascents upon which profiles are based, are shown under data. (Charnock, Francis, and Sheppard, 1956).

top scales,  $V$  (●) (m/s)



bottom scales,  $U$  (○) (m/s)

FIGURE 5 (cont.)



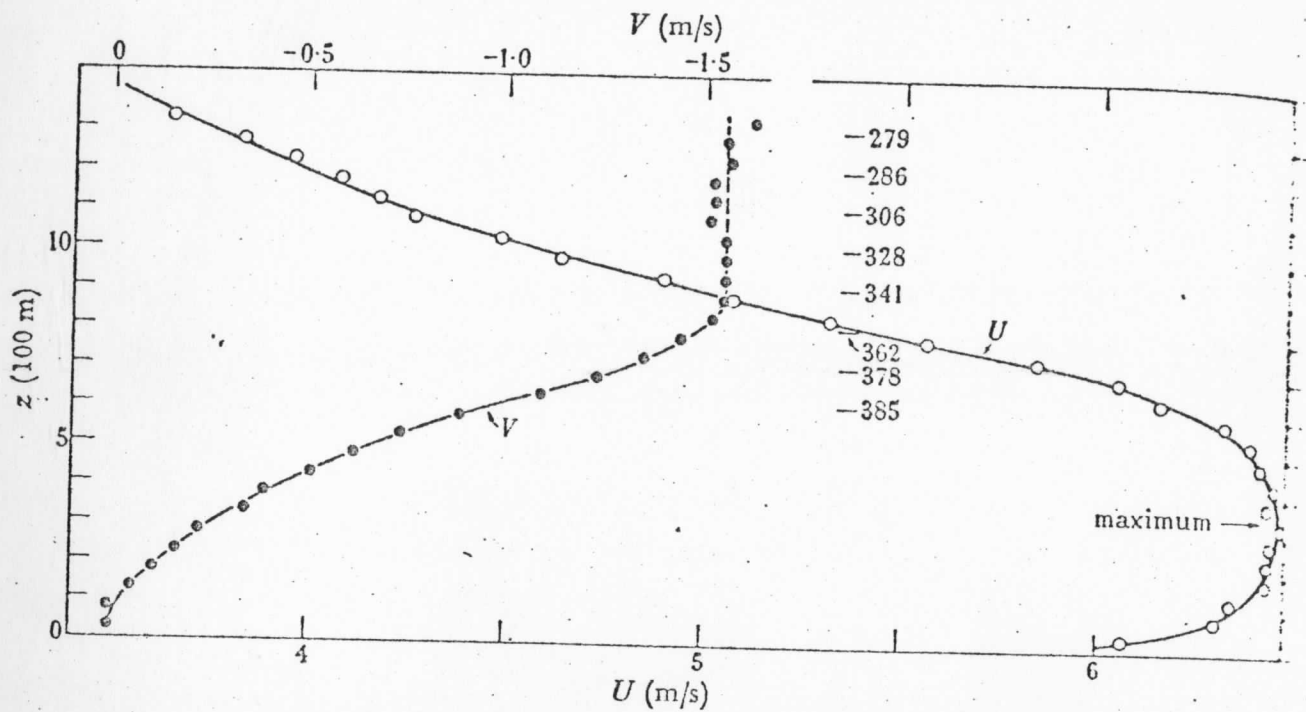


FIGURE 6a. The vertical profiles of the period-mean horizontal motion. Data for all 15 working days. Number of ascents to each level given above 600 m.

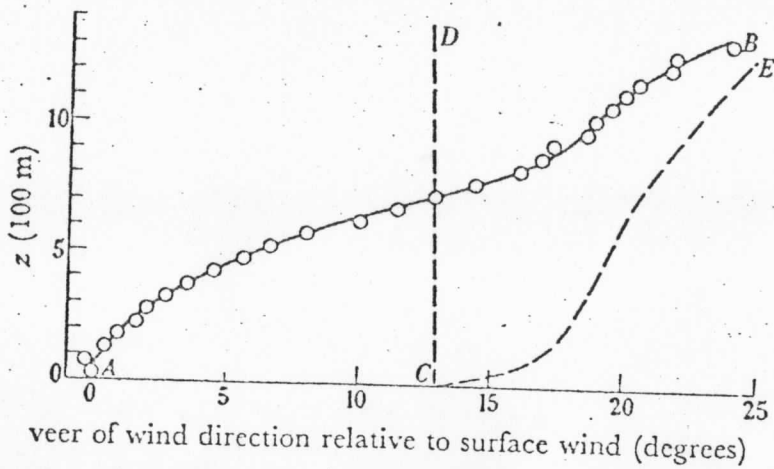


FIGURE 6b Period-mean profiles of observed and geostrophic wind directions. *AB* refers to the observed wind direction; *CD* to the geostrophic wind direction assuming the climatological temperature gradient; and *CE* to the best estimate geostrophic wind direction. (Charnock, Francis, and Sheppard, 1956).

Thermal wind contributions to profile variability are apparent in graphs constructed by Mendenhall (1967) for Johnston Island ( $17^{\circ}\text{N}$ ,  $170^{\circ}\text{W}$ ) in the Pacific and Swan Island ( $17^{\circ}\text{N}$ ,  $84^{\circ}\text{W}$ ) in Caribbean (Fig. 7).

The behavior of wind close to the equator was explored by Estoque (1971), using measurements made at Christmas Island ( $2^{\circ}\text{N}$ ,  $157^{\circ}\text{W}$ ) during the Line Islands Experiment. One to two day average profiles representing 17 to 24 individual soundings show a general similarity and relatively small vertical changes up to about 1500 m (Fig. 8). Above that level both zonal and meridional components become more variable. Estoque summarized these tentative results as follows:

1. Near the Equator, the change in wind direction (backing or veering) with height is controlled mainly by the height variation of the horizontal pressure gradient.
2. The ageostrophic wind component is of the same order of magnitude as the actual wind within a deep layer that extends from the surface up to the 1-km level.
3. The equation of motion in the east-west direction reduces to a balance between the horizontal pressure gradient force and friction.
4. The equation of motion in the north-south direction reduces to the geostrophic wind balance, except for a thin layer near the surface. Thus, the easterlies are approximately geostrophic.

The analysis preceding is intended as background for decisions on maximum sampling frequency consistent with experimental goals. Minimum sampling frequency is governed by the need to observe all impulses which may have a bearing on the development and evolution of cloud clusters. Recent work by Hoerber (1969, 1970) and Brier and Malkus (1969) suggests that the semi-diurnal pressure wave may be such an influence. Hoerber describes winds measured by a tethered buoy on the equator in the western Atlantic. The east-west component oscillated with an amplitude of 22 cm/sec (two per cent of the mean surface wind),

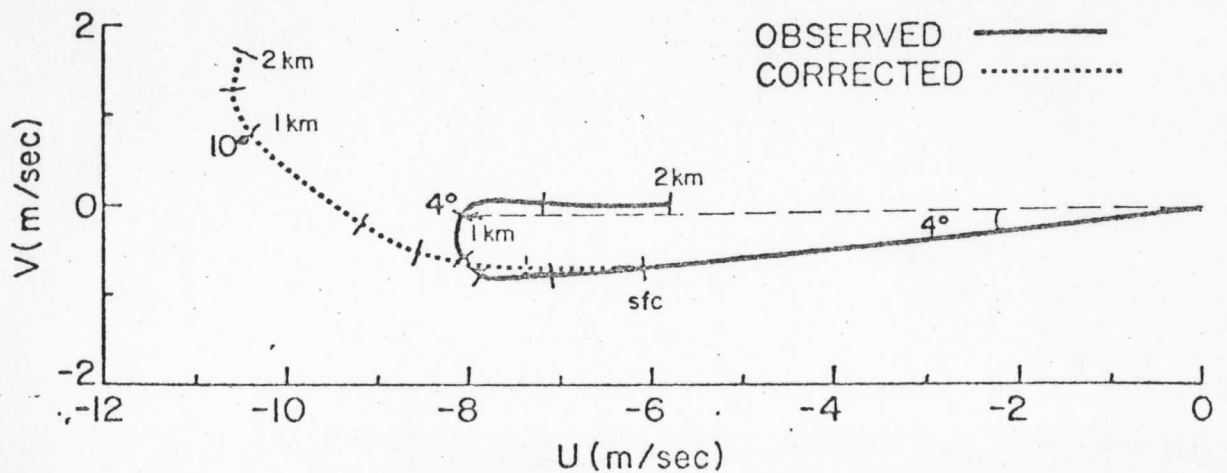


Fig. 7a. Observed and geostrophically corrected hodographs at Johnston Island derived from 3667 observations. Elevation marks are shown at the surface, 150, 300, 500, 1000, 1500, and 2000 m. Veering angle between the surface and 1000 m is noted on each hodograph. Observed veering is  $4^\circ$  and corrected veering,  $10^\circ$ .

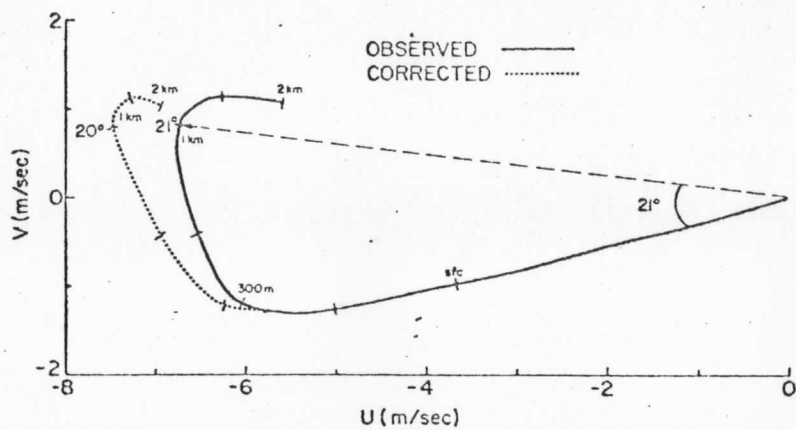


Fig. 7b. Observed and geostrophically corrected hodographs at Swan Island derived from 2070 observations. Elevation marks are shown at the surface, 150, 300, 500, 1000, 1500, and 2000 m. Veering angle between the surface and 1000 m is noted on each hodograph. Observed veering is  $21^\circ$  and corrected veering,  $20^\circ$ .

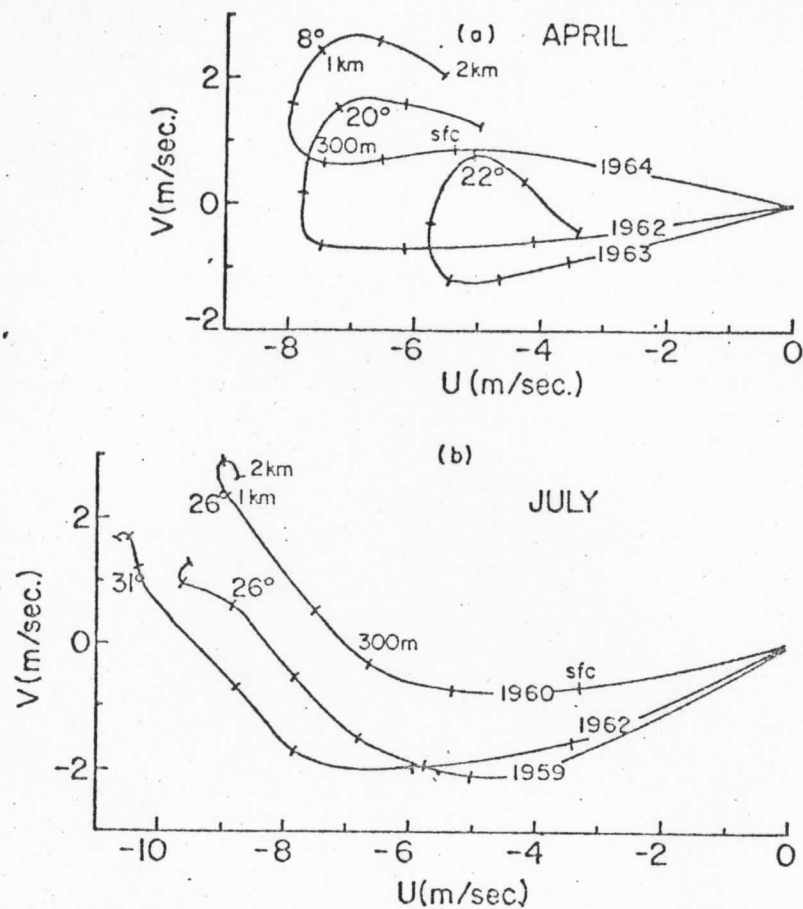


Fig. 7c. Typical (a) April and (b) July hodographs at Swan Island showing effect of reversal of meridional temperature gradient within the veering layer. Elevation marks are shown at the surface, 150, 300, 500, 1000, 1500, and 2000 m. Average veering between the surface and 1000 m is shown for each month. (Mendenhall, 1967).

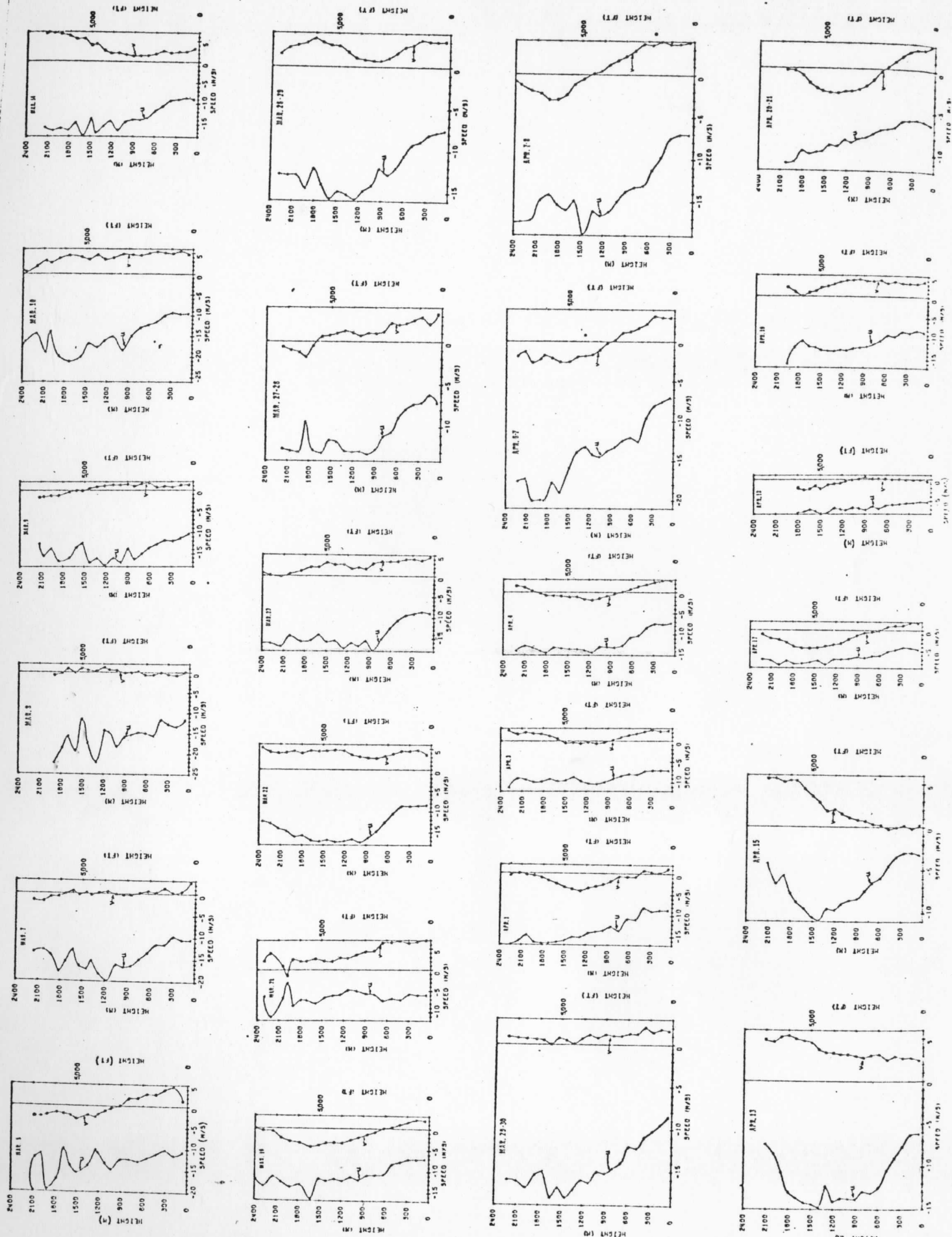


Fig. 8. Mean wind profiles for each of the observing periods (Estoque, 1971).

approximately in phase with the semi-diurnal pressure variation (Fig. 9). As Hoerber points out (1970), "A consequence of the semi-diurnal wave of zonal wind is a semi-diurnal convergence-divergence cycle within the trade wind region." Brier and Malkus (1969) demonstrated for Batavia (Djakarta) and Wake Island that large 5 to 6 hourly pressure changes of the semidiurnal solar pressure ( $S_2$ ) wave were accompanied by large cloudiness changes relative to days of small pressure changes. They concluded that "the  $S_2$  effect acts to increase cloudiness and rain near sunrise and sunset and to suppress them near midday and just after midnight."

The spectra presented by Estoque (1971) and an analysis of a Pacific equatorial disturbance by Zipser (1969) suggest that active clusters are associated with strongly disturbed flow in the boundary layer. It is important to the operation of a tethered balloon system to know how long these disturbed conditions can be expected to last.

At Fanning Island disturbed winds associated with the small system of Zipser's study persisted for about six hours (Fig. 10). No comparable measurements are presently available for the Atlantic; however, some idea of disturbance duration may be gained from consideration of the mean size and speed of Atlantic disturbances as deduced from satellite pictures. A mean cloud cluster size of  $5 \times 10^5 \text{ km}^2$  has been found by Martin (unpublished). Assuming cloud area corresponds approximately to the region of strongly disturbed winds, a movement of 8 m/sec (the average found by Simpson, Frank, Shideler, and Johnson (1969) for Atlantic disturbances of 1968) would result in 24 hours of disturbed conditions at a fixed station crossed by the disturbance. This time could easily be doubled by the combination of a larger system and a slower movement.

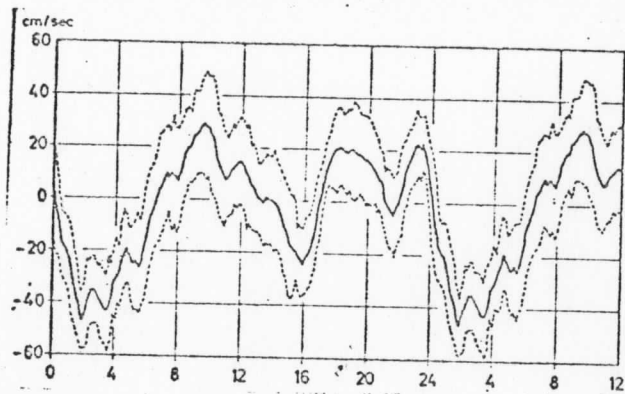


Fig. 9. Mean diurnal course of wind speed at 4 m for the period of Sept. 21 through Oct. 9, 1965. Dashed curves give probable error of the mean ( $n \approx 16$ ). (Hoerber, 1970).

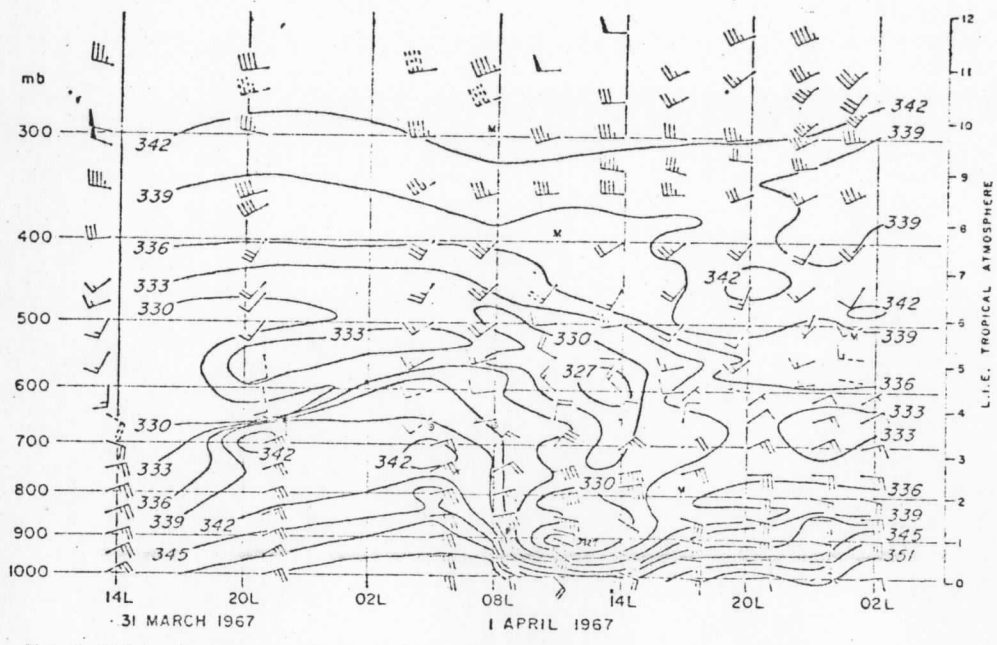


FIG.10. Height-time cross-section at Palmyra. The rawinsonde winds are plotted in the conventional manner, with each full barb representing 5 m sec<sup>-1</sup>. Dashed wind arrows represent reduced confidence in that observation. The analyzed scalar field is equivalent potential temperature (°K) computed from the temperature and humidity data on the same soundings. The leading edge of the downdraft air passes Palmyra at 0800. (Zipser, 1969).



## B. Depth

The figures preceding illustrate the difficulty of defining "boundary layer" in a way that has meaning over equatorial as well as trade latitudes. "Ekman layer" fails near the equator; "friction layer" is excessively restrictive. For purposes of the GATE, it may therefore be best to extend observations to the base of the trade inversion, that is, through the layer of nearly constant potential temperature and slowly decreasing mixing ratio. This approach has the advantage of providing a substantial, extensive, and meteorologically significant upper bound on the layer to be sampled.

The charts of Ficker, reproduced in Weather on the West Coast of Tropical Africa (1949), show that over most of the B scale area selected for GATE the base of the trade inversion is found below 1500 m (Fig. 11). To the extent that comparisons are possible, this is supported by measurements of the ATEX drift experiment during February 1969 in the central Atlantic, as reported by Brocks (1970) (Fig. 12).

## C. Accuracy of Measurement

The need for extremely accurate measurements of pressure is apparent from examination of a "typical" surface pressure pattern, constructed by adding a disturbance field to the mean field for July and August (Fig. 13), which is given in Weather on the West Coast of Tropical Africa (1949). The disturbance field is a space extrapolation of pressure oscillations (with a 3 mb peak to peak amplitude taken to be representative) at Praia (15°N 24°W) in the Cape Verde Islands, as reported by Carlson (1969). Assumed movement of the pressure disturbance is westward at 8 m/sec. The disturbance center is positioned just east of the area specified for location of the B-scale ship network. Profiles of pressure along the 12° latitude line show for two

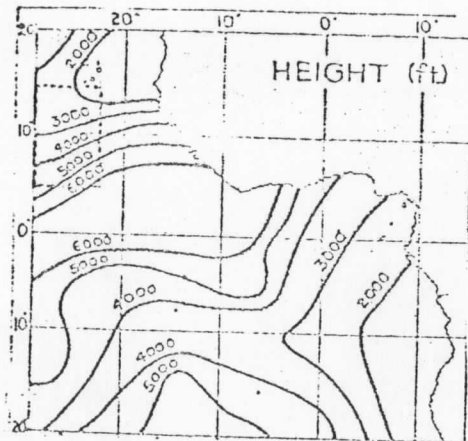


Fig. 11. The average height of the base of the inversion, referenced mainly to the northern summer. The area recommended by JOC V for the B-scale experiment is added. (Weather on the West Coast of Tropical Africa, 1949).

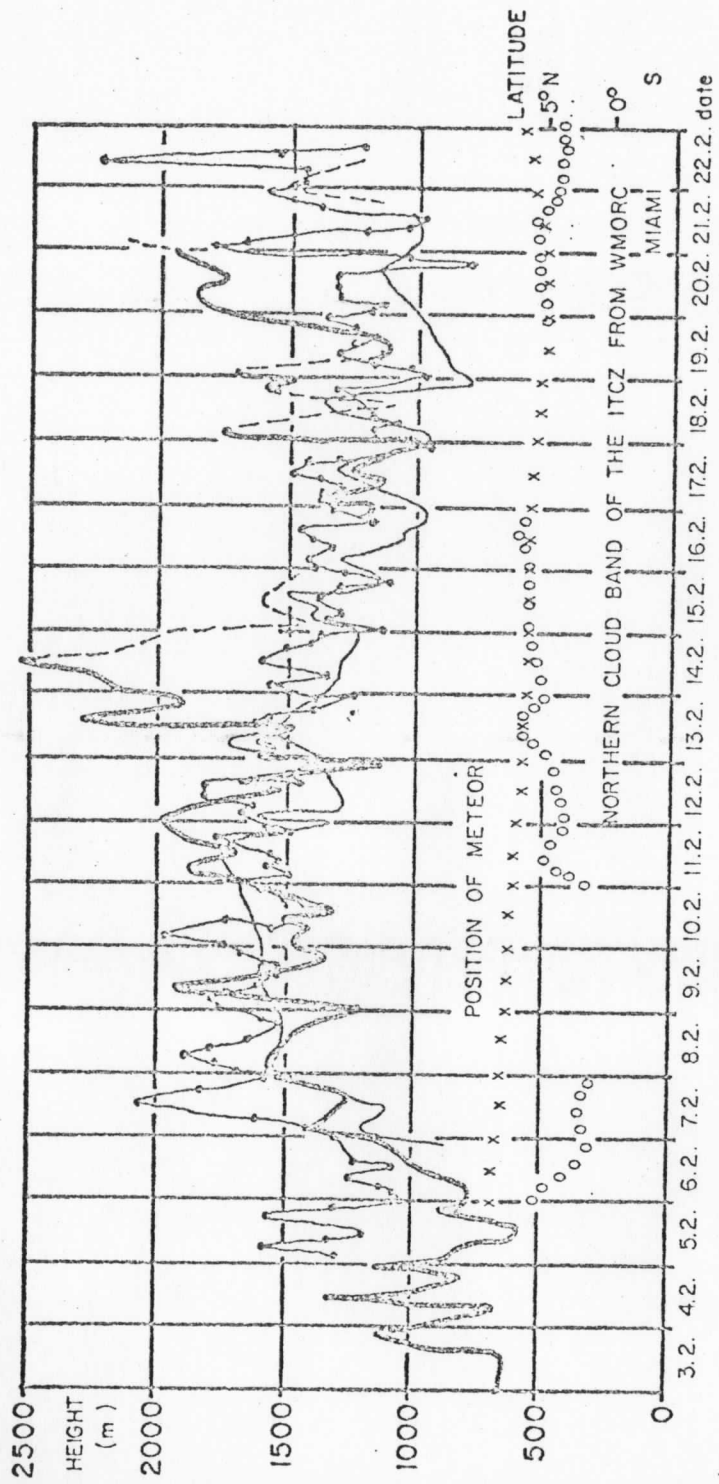
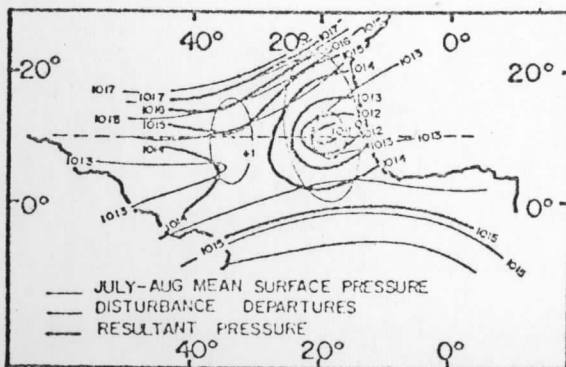
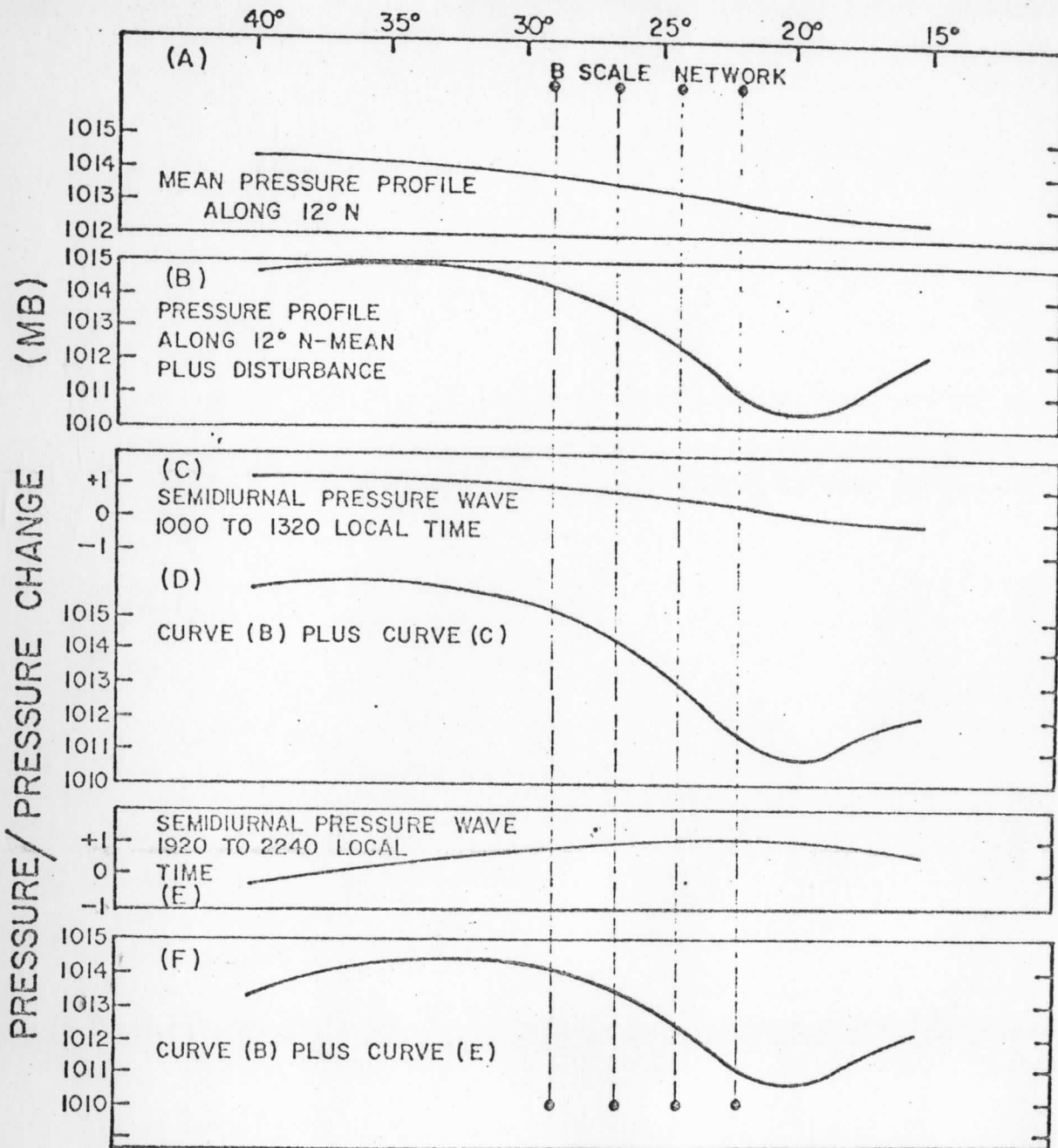


FIG.12. HEIGHT VARIATION TRADE WIND INVERSION DURING ATEX 1969

METEOR ——— PLANET ——— DISCOVERER ———

MET. INST. HBG.

# LONGITUDE



## PRESSURE GRADIENTS ALONG THE 12° LATITUDE LINE

CURVE	1-2	2-3	3-4
(D)	0.8	1.3	1.8 mb/260Km
(F)	0.6	0.9	1.4 mb/260Km
(A)	0.2	0.2	0.2 mb/260Km

Fig. 13. Pressure profiles for a hypothetical surface pressure field--disturbance plus mean--over the eastern Atlantic.

configurations of the semidiurnal pressure wave the size of gradients between ships of the B-scale network. In the most extreme case (curve d) pressure differences approach 2 mb. A more typical value is 1 mb, and away from the region of largest gradient, differences for stations of this spacing are on the order of 0.5 mb.

Errors in the measurement of pressure gradients are larger for elevated pressure surfaces. The example presented in Table 1 shows errors involved in a hypothetical computation of the height of the 850 mb surface. For an absolute dry bulb temperature error of  $\pm 0.1^\circ\text{C}$ , a relative humidity error of 2%, and a pressure error of  $\pm 0.1$  mb the height error is  $\pm 2.7$  m. This is equivalent to a 0.5 mb error in the measurement of the horizontal pressure gradient, which is only slightly smaller than typical gradients (assuming no change of intensity up to 850 mb) inferred from Fig. 13 over a 260 km distance.

Davies-Jones and Ward (1971) argue that within cumulonimbus updrafts corrections for dynamic pressure, entrainment, and liquid water content must be made to thicknesses computed from the hypsometric equation. The corrections

are of the form

$$z_2 - z_1 = \frac{R}{g} \int_{P_1}^{P_2} T_v d(\ln p) - \frac{1}{2g} (w_2^2 - w_1^2) - \frac{\mu}{g} \int_{z_1}^{z_2} w^2 dz - \int_{z_1}^{z_2} \sigma dz$$

(1)
(2)
(3)

where  $Z$  is height,  $R$  is the gas constant for dry air,  $g$  is acceleration due to gravity,  $p$  is pressure,  $T_v$  is virtual temperature,  $w$  is vertical velocity,  $\mu$  is a constant entrainment coefficient, and  $\sigma$  is liquid water mixing ratio. Taking a simple example, with the following values assumed for the independent variables-- $g = 10 \text{ m/sec}^2$ ,  $w_2 = 5 \text{ m/sec}$ ,  $w_1 = 0 \text{ m/sec}$ ,  $\mu = 2 \times 10^{-4}/\text{m}$ ,  $\bar{w} = 2 \text{ m/sec}$ ,  $\sigma = 4 \times 10^{-3} \text{ gm/gm}$ , and  $z_2 - z_1 = 1400 \text{ m}$ --the values of the terms (1), (2), and (3) total 7 m. Such an error could completely mask real

gradients. Therefore, if the slopes of pressure surfaces in the upper part of the boundary layer are to be accurately measured under disturbed conditions, it is imperative that geometric altitude be independently measured with an accuracy of at least  $\pm 2$  m.

Williams (1970) has found that divergence values of  $\sim 5 \times 10^{-6} \text{ sec}^{-1}$  are associated with cloud clusters in the tropical Pacific. For a ship spacing of 250 km measurement of such a divergence to an accuracy of 10% requires measurement of winds with an accuracy of 0.05 to 0.1 m/sec, that is, 1% of wind speeds typical of the tropical trades and one degree of wind direction.

Temperature differences between disturbed and undisturbed regions are reported to be 1 to 2°C at the surface (Wallace, 1971); therefore, measurement accuracies of  $\pm 0.1^\circ\text{C}$  should be sufficient to define the disturbance thermal field.

Since the variation of relative humidity, as reported by Wallace (1971), increases from nil in the subcloud layer to 10 to 30% above the subcloud layer, two per cent absolute accuracy of measurement should be sufficient to define larger fields of anomaly above the subcloud layer.

### III. Sampling Requirements

#### A. Turbulent Time Scale Variations

The tropical atmosphere within the boundary layer has the following characteristics of variability:

Below the cloud level

1. Time variability on the turbulent scale on the order of seconds but less than minutes.
2. Time variability on the order of tens of minutes to a fraction of an hour due to organization of the flow into rolls, lines, gravity waves, etc.
3. A time variability on the scale of hours due to the effects of the semi-diurnal atmospheric tide and due also to the passage of disturbed weather.
4. A time variability of days due to synoptic scale motions.

Within the Cloud Layer

While hard data on the variability in this layer of the atmosphere is almost non-existent, one can estimate with fair confidence that the variability in class A (1) above ought to be reduced because in the absence of strong vertical shear there is no mechanism to develop the fine scale turbulence. On the other hand, variability of the type described in A (2) above ought to be greater due to the release of heat of condensation in clouds on time scales commensurate with the passage of cloud elements. The longer scales of motion should be similar to those in the regions below the cloud layer.

In sampling the atmosphere one must obey Shannon's sampling theorem which states: "A signal whose highest frequency component is  $f$  is completely determined by a sampling scheme whose sampling rate is  $2f$ ."

One must therefore sample at a rate twice as high as the highest frequency in the signal developed by the sensor. A sensor with a long time constant acts as an integrator of the rapid signal fluctuations and is an easy way to limit the frequency composition of the signal from the sensor. Hall (1949) has shown the frequency composition of a signal having a rise time

in the shape of a hyperbolic tangent (instead of an unrealistic step function), and a pulse in the shape of a probability function (instead of an unrealistic box) is a strong function of the time constant. Figures 14, 15, and 16 from his paper show the frequency composition of the signal at various time constants of the sensor. Clearly, high frequency variability of an atmospheric phenomenon can be reduced significantly by the use of a sensor having a long time constant. Thus the high frequency variability in the signal due to the passage of turbulent elements can be suppressed to a suitable level if we have thermometers and anemometers which have slow speeds of response. Stated simply, the performance of a tiny bead thermistor temperature element will be improved if it is imbedded in a wad of chewing gum! Alternately, one can sample at a high rate and do the integration later in the data processing. What one cannot tolerate is a sensor having a high speed of response and a data scheme with infrequent sampling, because this leads to very serious aliasing errors.

Variabilities having time scales of several hours or days do not constitute a difficult measurement problem other than meeting the requirement of at least 4 samples per day in order to properly sample the atmospheric changes tied to the semi-diurnal atmospheric tides.

The most difficult sampling problem is that which arises from time variability on the order of tens of minutes to an hour. It should be clearly understood that if one conducts a series of measurements often enough to meet the requirements of the turbulent scale he has met the requirements for the tens of minutes to an hour scale. We have a problem because in addition to obtaining adequate time resolution we must also obtain adequate vertical space resolution.



$G(t)$	$G(t)$	$S(\omega)$	$S(\omega)$
(a) Unit step $G(t) = +1$ (from 0 to $\infty$ ) $G(t) = 0$ (from $-\infty$ to 0)		$\frac{1}{\omega}$	
(b) Single square wave of duration T $G(t) = 0$ (from $-\infty$ to 0) $= +1$ (from 0 to T) $= 0$ (from T to $+\infty$ )		$\frac{2}{\omega} \sin\left(\frac{\omega T}{2}\right)$	
(c) Single impulse		Constant	
(d) Hyperbolic tangent $G(t) = \frac{1}{2} \tanh at$		$\frac{\pi}{2a \sinh\left(\frac{\pi\omega}{2a}\right)}$	
(e) Probability function $G(t) = e^{-at^2}$		$\sqrt{\frac{\pi}{a}} e^{-\frac{\omega^2}{4a}}$	
(f) Oscillation of frequency $\omega_r$ lasting from $t = -\infty$ to $t = +\infty$		$\omega_r$	

Fig.4. Elementary signals and their frequency composition.

— (Hall, 1949).

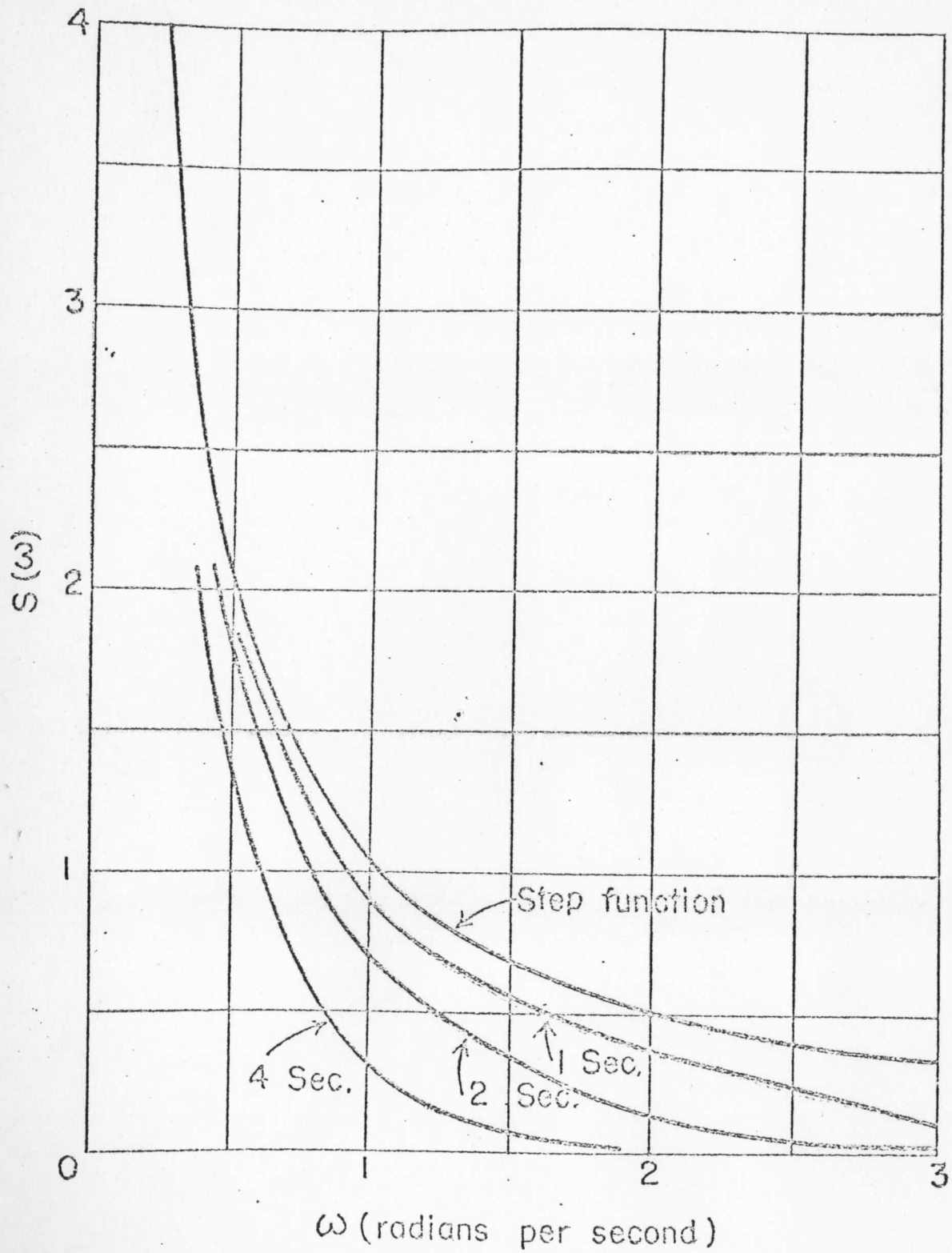


Fig. 15. Frequency composition of step function, and of hyperbolic tangent for "rise-times" of 1, 2, and 4 seconds. (Hall, 1949).

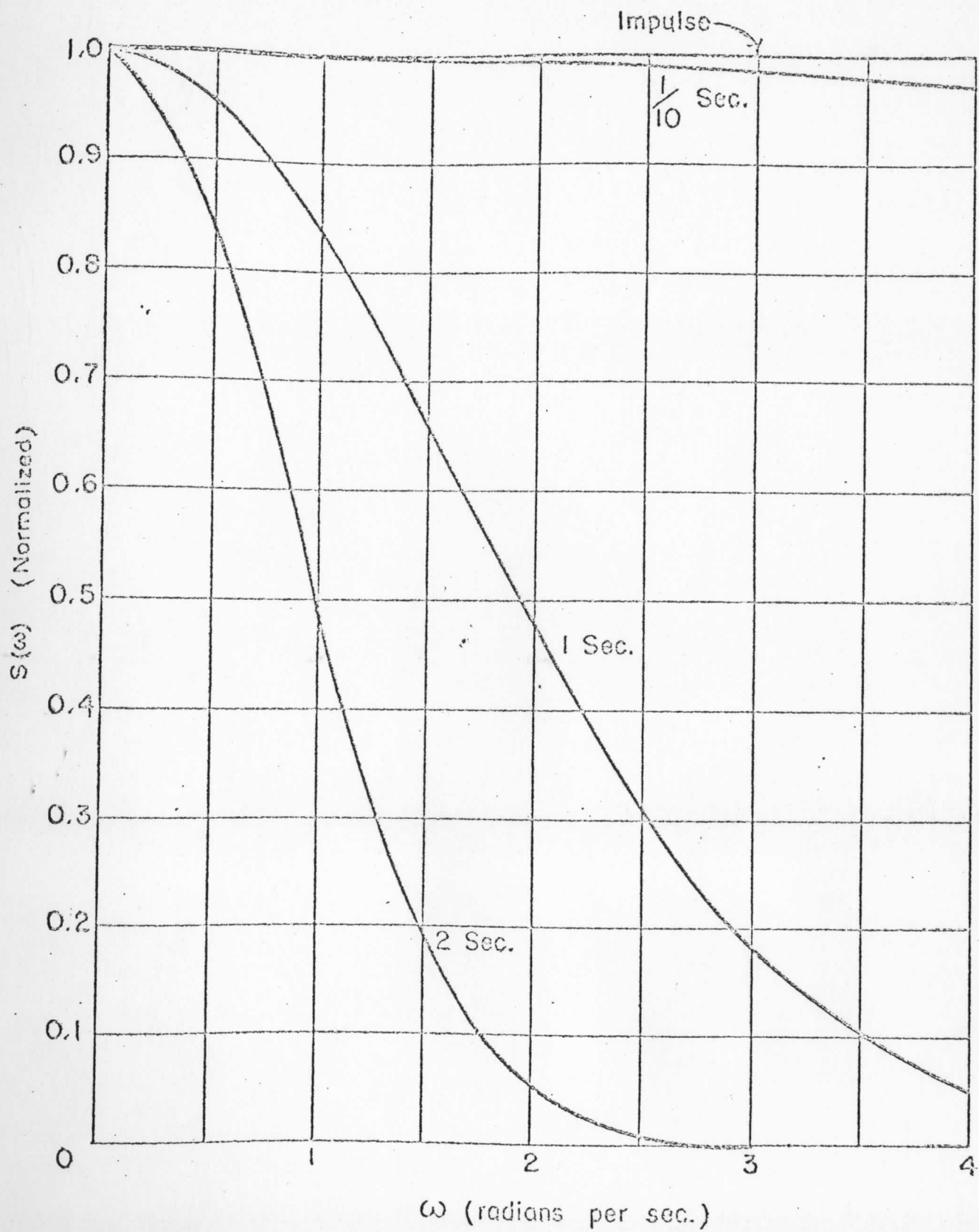


Fig. 16. Frequency composition of impulse, and of probability function for "rise-times" of 1/10, 1, and 2 seconds. (Hall, 1949).

## B. Vertical Variations

Figures 4 to 8, taken from a variety of observers, show large variations of the boundary layer wind with height, despite the smoothing which is inherent in averages of many individual soundings. They illustrate the difficulties of obtaining complete sampling in the vertical with a limited number of fixed altitude observations.

We are faced with a dilemma: In order to properly sample the vertical fine structure we need a sounding, i.e. a balloon ascent; however, in order to properly sound the horizontal, we must sample long enough to remove the time variability on the order of minutes. A further complication is the need to sample often enough to capture the semi-diurnal oscillation. Ideally, one would wish for many ascents several times an hour. Operationally, this is not practical.

A high correlation between simultaneous measurements at adjacent levels would simplify the sampling problem, as is shown by the equation relating the variance of the difference of the means at two levels to the sum of the variances of the means minus the covariance of the means, i.e.

$$\sigma_{\bar{x}-\bar{y}}^2 = \frac{\sigma_x^2}{n_x} + \frac{\sigma_y^2}{n_y} - 2r \frac{\sigma_x \sigma_y}{\sqrt{n_x n_y}}$$

Assuming  $\sigma_x$  and  $\sigma_y$  are approximately equal (as they are likely to be), a high correlation coefficient implies near zero variance for the difference of the means. Once the difference is specified one can measure the mean at each level quite separately, or he could slowly probe the atmosphere in the vertical and be assured of excellent results.

Some insight as to the extent of correlation in the tropical boundary layer is to be found in the observations of Figures 4 through 8. The presence of horizontal rolls and waves implies a correlation, as does friction, the vertical transport of horizontal momentum. However, the layers of strong frictional influence may at times be quite shallow, and the horizontal circulations of rolls are intermittent. Since the BLIS must be effective over all conditions short of hurricane, vertical correlation is inadequate as a basis for sampling design, but whatever correlation exists will improve the accuracy of the means.

#### IV. Systems

The analysis preceding, considered in the context of experimental conditions, is sufficient to eliminate from consideration many observational systems used in the past. Among these are double theodolite, rawinsonde, smoke puffs, chaff, and rockets, all of which are incapable of providing at reasonable expense the near-continuous, time-averaged, extremely accurate values which are required. Free floating balloons under existing navigation aides cannot return required accuracies of wind, and pose an enormous operational and analysis problem. Parafoil kites require wind, which, even in the trades, is not always present. The tethered balloon carries its own handicaps, but has the great advantage of providing continuous observations over time periods approaching a day. It is the platform which is best suited for general observations of the tropical boundary layer during GATE. The remainder of this report will consider various configurations of instrument packages on a tethered balloon.

The simplest configuration, that which was used during BOMEX, has a single instrument package clamped to the tether. Package and balloon ascend and descend together. This system is capable of returning profile or fixed level measurements, but is limited to one measurement level at any given time.

Multiple packages clamped to the tether at regular intervals return simultaneous observations at several levels; however, each package added decreases free lift and increases operational complexity. Cycling the system through steps of ascent and descent improves vertical sampling density, but adds still further to operational complexity.

As an alternative to winching both balloon and package to obtain profiles, the instrument package can be clamped to a separate line run around a pulley at the base of the balloon. This configuration puts additional weight on the balloon, requires a second winch, and adds the hazard of tangled lines. Further, the balloon itself must be winched in regularly for topping off.

The objection to profiles is that time averaging is required to eliminate small scale fluctuations. The profile must therefore be considered as a series of vertical averages, the interval of the average depending on speed of ascent or descent. If it is assumed that averages of one hour are necessary to eliminate aliasing due to lines and rolls the requirement that semi-diurnal changes be observed limits sampling intervals rather severely, to three for each half cycle. Averaging over 30 minutes would increase this number to six.

A package that holds at predetermined levels combines profile and constant level sampling. Allowing 30 minutes for winching through a cycle of ascent and descent (equivalent to a winching speed of 2 m/sec over a 1500 m sampling interval), the maximum number of levels that could be sampled and averaged over an hour is three. Averaging over 30 minutes would increase this number to five levels. These sampling models are depicted in Fig. 17.

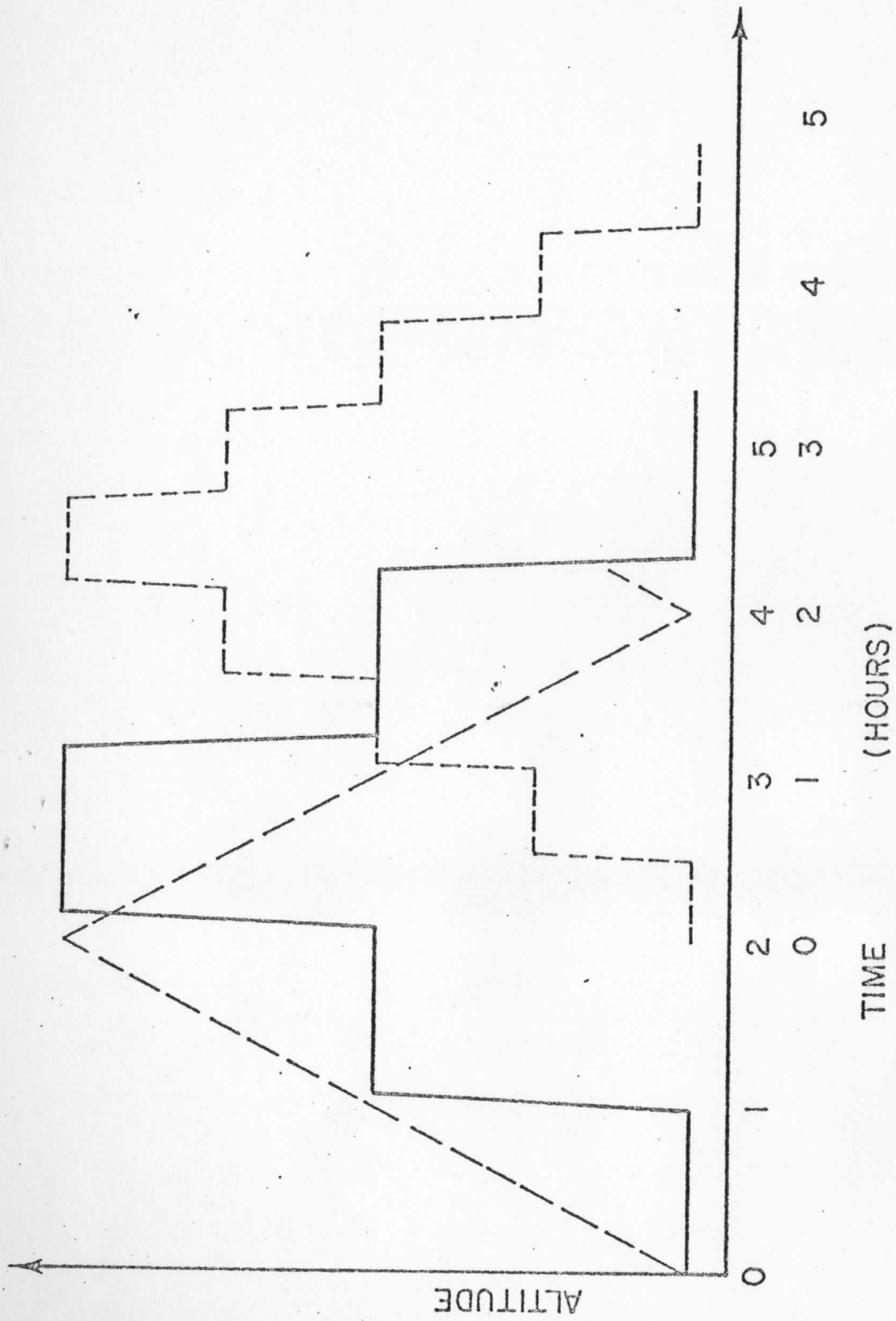


Fig. 17. Single package sampling schemes.

Multiple platforms are limited by the balance between balloon buoyancy and package weight. Achievement of a target weight of 0.5 kg for each BLIP allows for one Jalbert J-8 balloon (the type used during BOMEX) a total of 5 to 7 packages spread over 1500 meters of altitude. If one hour averages are needed to eliminate the effects of roll and line organization, multiple platforms have a definite edge. Yet even this number will at times fail to adequately portray the vertical structure of the boundary layer.

Better definition of the vertical structure can be attained through use of balloons having greater buoyancy, or the combination of multiple platforms with the sonde technique in a system which cycles several platforms in vertical steps. The step may simply be equal to the distance between platforms (assumed equal in this discussion) in which case the sampling gain is one level plus profiles between levels; or the step may be some fraction of the distance between platforms (Fig. 18). If the step is one half the platform interval; i.e., two steps down and two steps up, the gain in levels is  $n + 1$ , (total  $2n + 1$ ) where  $n$  is the number of platforms. Thus five platforms holding at each level for one hour could return time averaged values at eleven levels (twenty values altogether) in just over four hours.

The multiple platform, stepped level configuration is analyzed by considering variations on a nominal case (Fig. 19). A balloon supports four equally spaced instrument packages, which are attached to a tether inclined at an angle  $\delta$  to the local vertical. The balloon and instruments ascend and descend in a stepwise pattern, so that all but the lowest 100 meters of the boundary layer is sampled. It is assumed that each step is some fraction of the package interval. Steps, hold times, and rate of ascent and descent are equal through the sampling cycle.



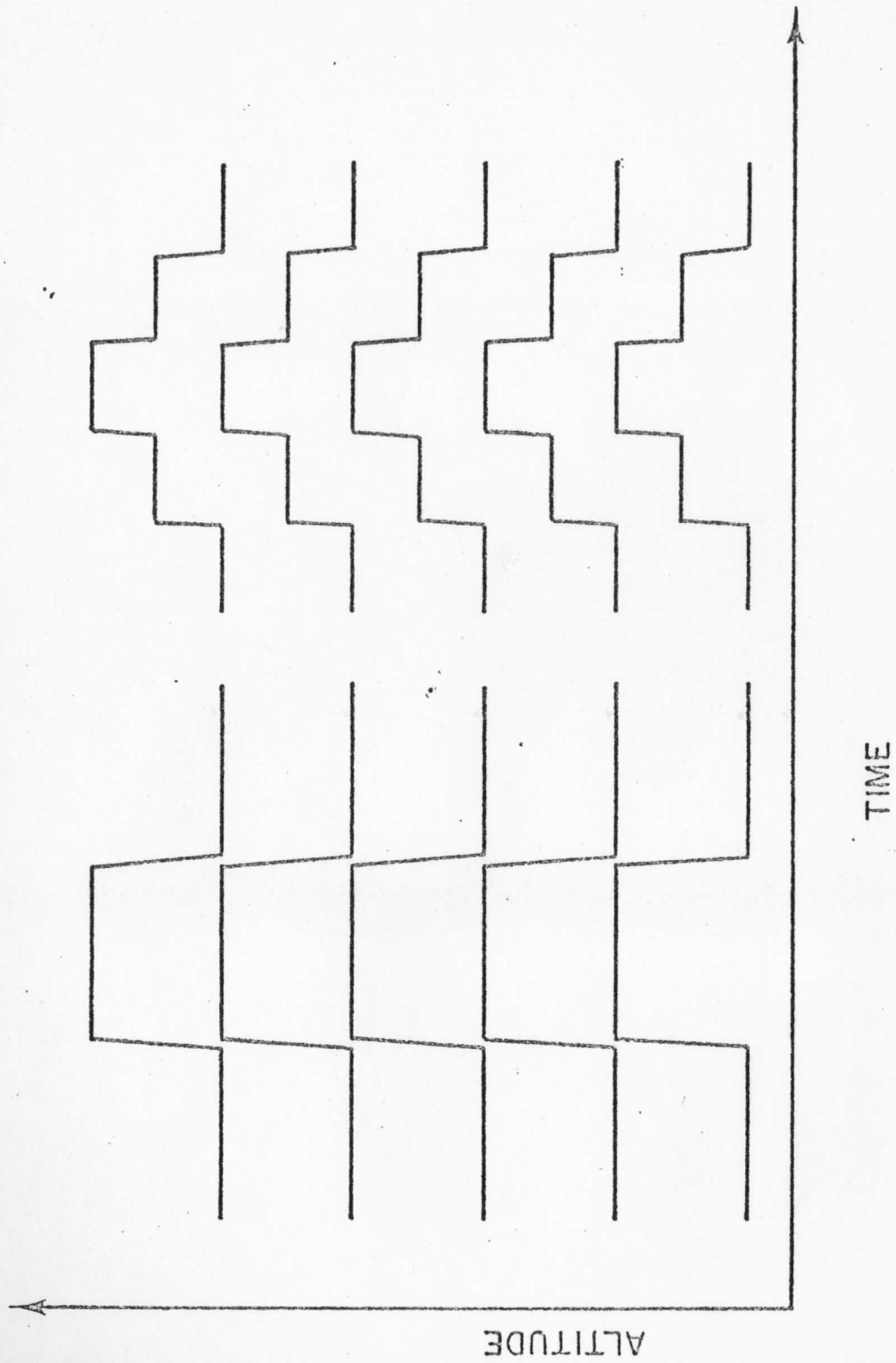


Fig. 18. Multiple package, stepped level sampling schemes.

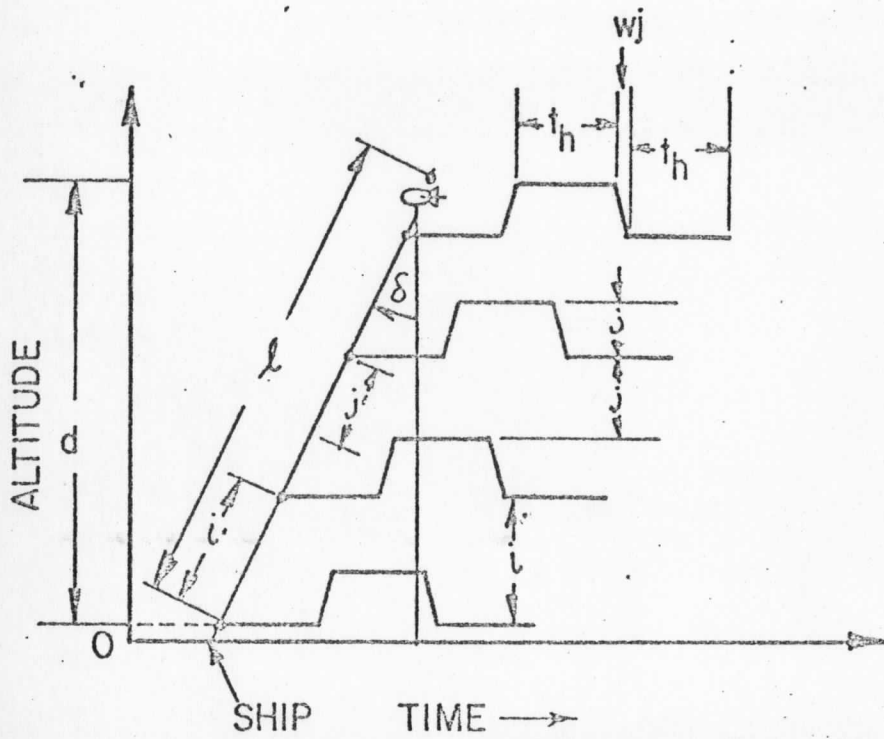


Fig. 19. Example of a multiple platform, stepped level sampling scheme.

Symbols used are described below.

$d$  = thickness of the layer to be sampled

$\delta$  = angle between tether and the local vertical

$l$  = length of tether through  $d$ ;  $l = d/\cos \delta$

$w$  = rate of change of platform height

$n$  = number of BLIP's

$s$  = number of steps per cycle

$i$  = vertical interval between BLIP's

$i'$  = tether distance between BLIP's;  $i' = i/\cos \delta$

$j$  = vertical interval between step levels

$j'$  = tether distance between step levels;  $j' = j/\cos \delta$

$k$  = some fraction relating  $j$  to  $i$ ;  $j = ki$

$t_h$  = hold time at each level

We let  $t_c$  be the time required for one cycle. Then

$$t_c = \sum \text{step times} + \sum \text{hold times}$$

$$= \frac{s}{w} \left[ \frac{d/\cos \delta}{s/2 + (n-1)/k} \right] + st_h$$

Nominal values assumed for the independent variables in this equation are

$$s = 2$$

$$n = 4$$

$$w = 120 \text{ m/min}$$

$$k = \frac{1}{2}$$

$$d = 1400 \text{ m}$$

$$t_h = 30 \text{ min}$$

$$\delta = 5.4^\circ$$

The influence of each variable on the cycle time  $t_c$  is shown in Fig. 20.

Part (a) considers the effect of increasing distance due to tether inclination.

This, for a rate of change of 120 m/min, is very small, and would remain small

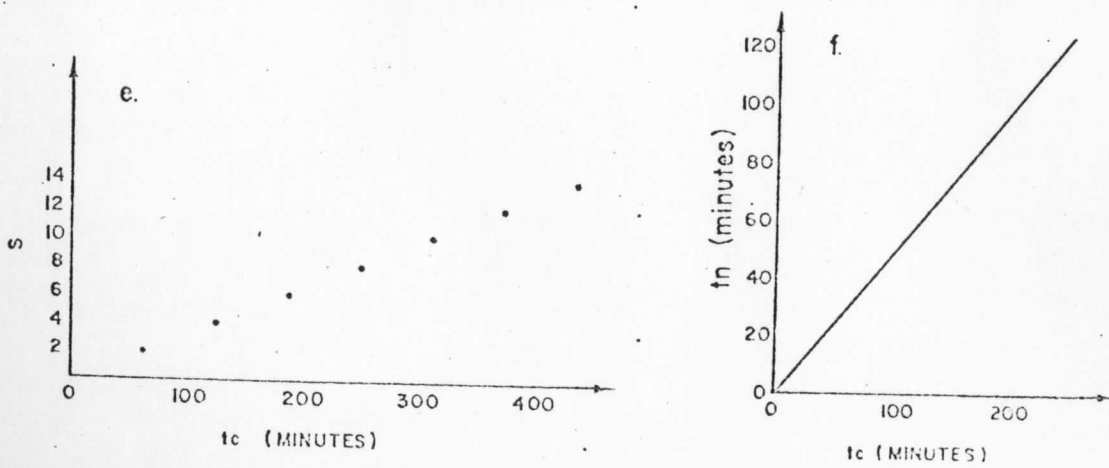
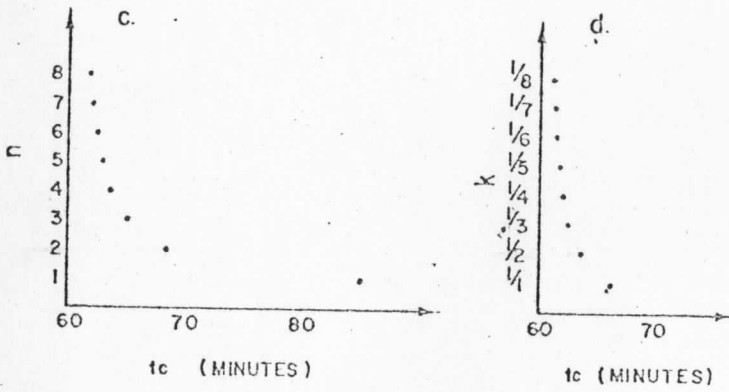
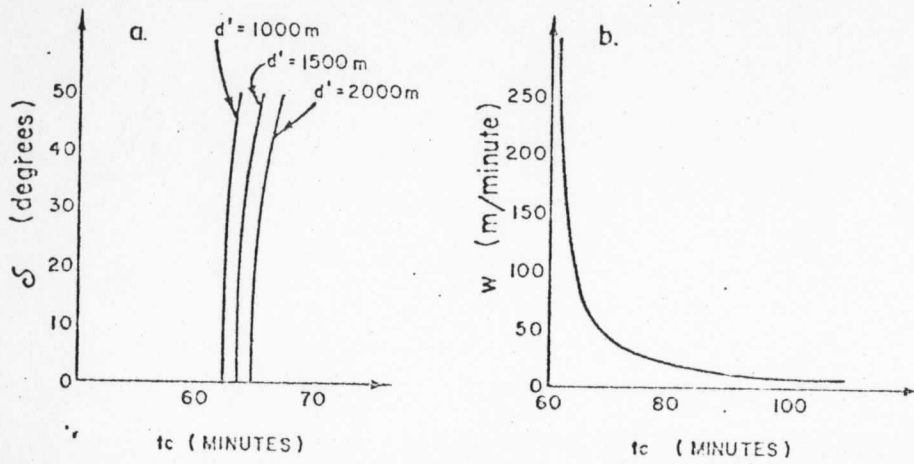


Fig. 20. Analysis of Variables Affecting Total Cycle Time in a Multiple Platform Stepped Level Sampling Scheme.

down to rates a tenth of that assumed. Part (a) also shows the insensitivity of total cycle time to changing boundary layer thickness. Rates of level change greater than 50 m/min (Part b) assure small contribution to  $t_c$  of the step time. Increasing the number of packages beyond 3 (Part c) gives little additional reduction in  $t_c$ . For any step interval less than the package interval (Part d) changing  $k$  has little effect on  $t_c$ . The dominant effects on  $t_c$  are the number of steps (Part e) and the hold time (Part f). A small increase in either will substantially lengthen the cycle time.

A problem with the multiple platform, stepped level scheme is the operational complexity of the stepped cycle; however, this type of sampling could be a special mode, for use only when weather systems of particular interest are within the sampling network. Offsetting operational complexity is the advantage of simple calibration of each instrument package against the package above or below.

#### V. Conclusions and Recommendations

A boundary layer system capable of meeting the requirements of the GARP Atlantic Tropical Experiment must be able to

1. Operate on a near continuous basis, under all but hurricane conditions.
2. Measure to at least 1000 m altitude; 1500 m would be far better.
3. Return winds of sufficient accuracy to infer cluster scale divergence and convergence.
4. Define vertical structure in detail sufficient for computation of total boundary layer convergence and divergence.
5. Withstand the abuses and tolerate the limitations of operation in a confined space on a pitching, rolling platform by men who in the stress of these conditions may have little skill and less motivation.

We conclude that the system best able to meet these requirements consists of multiple instrument packages attached to the tether of a helium airfoil balloon. Under normal operation these packages will be maintained at fixed levels. When conditions require greater vertical resolution the balloon and its packages will be lowered and raised in a stepwise cycle.

TABLE 1: ERRORS IN HYPSONETRIC HEIGHT COMPUTATIONS

<u>PARAMETER</u>	<u>ERROR</u>	<u>EQUIVALENT HEIGHT ERROR</u>
1000 mb (surface) pressure	<u>+0.1</u> mb	<u>+0.9</u> m
850 mb pressure	<u>+0.1</u> mb	<u>+1.0</u> m
mean virtual temperature 1000 to 850 mb	<u>+0.2</u> °C	<u>+0.8</u> m
total		<u>+2.7</u> m

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