The Effect of Radiation Absorption in the Tropical Troposhere

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ABSTRACT

Data from an experiment to measure the upward and downward components of solar radiation from aircraft during the Barbados Oceanographic and Meteorological Experiment (BOMEX) have been analyzed in the present study. Results show that in the cloud free tropical troposphere: 1) Absorption of solar radiation in the entire troposphere can be twice as large as previous estimates given by Manabe and Strickler(13% absorption; 1964) (Comparison of observed heating rates to calculations shows that the increase in attenuation may be due to non-gaseous constituents in the atmosphere); 2) The vertical profile of solar radiative heating was particularily variable in the lowest layers; in the mean, solar heating showed a slight maximum near 700 mb; and 3) In using the solar radiation observations of this study in an energy budget of the tropics, a hypothesis regarding a nighttime maximum of preciptation in the tropical regions is formed.

Results from a radiative study of certain cloud type cases show that: 1) Selective vertical absorption in strato-cumulus acts to destabilize the clouds' environment; 2) Large cumulus, or cumulonimbus clouds occasionally decrease the solar insolation reaching the surface to only 3% of that incoming at cloud top; and 3) cirrus clouds are possible stabilizing mechanisms in the tropical environment since they act to warm the local environment at high altitudes while suppressing solar warming from cloud base to surface.

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1.0 INTRODUCTION

It is known that the absorption of solar energy is a heat source to the atmosphere with its role in atmospheric energetics varying with location, season, time of day and altitude. This study will deal specifically with measurements of absorption in the tropical atmosphere from approximately 0.3 km to 12 km. A detailed study of the solar radiation characteristics at the surface will not be considered here.

In the tropics, where a large amount of solar insolution occurs each day, horizontal variation of solar radiation absorption can provide a rapid energy input differential to drive atmospheric motions. In addition, selective vertical absorption can affect local atmospheric conditions which may result in an increase or decrease in tropical weather activity, i.e., cloudiness or rainfall.

Only in the last 10-15 years have there been reports of measurements made of solar absorption in the tropics. These measurements have been acquired from instrumented aircraft making vertical soundings in the atmosphere. Roach (1961a) carried out measurements in 1958 and 1959 over Southern England, the Sahara Desert and Equatorial East Africa using a Canberra Aircraft and found that in the clear atmosphere, under hazy conditions, instantaneous heating rates in excess of 5° C/day occur quite frequently. Griggs (1968) used a single aircraft to measure the albedo and absorption of stratus clouds off the coast of Southern California. Paltridge (1971) conducted aircraft measurements over the ocean near Australia measuring the radiative characteristics of trade wind stratocumulus clouds as well as clear areas in which dust from burning sugar cane fields had been picked up by the wind and suspended in the air. Other measurements of solar radiation absorption have been made by using specially instrumented radiosonde packages capable of measuring the upward and downward flux of solar radiation (Kondratyev, 1972b; Queck et al., 1972). These measurements pointed out the strong absorption taking place by atmospheric aerosols. Fritz et al. (1964); Hanson et al. (1967); and Vonder Haar and Hanson (1969a) all have used both ground based solar radiation measurements and satellite reflected energy measurements to derive values of absorption for an entire atmospheric column.

Many of the radiation measurements have been made in an effort to improve upon theoretical models dealing with solar energy absorption and its heating of the atmosphere. Yamamoto (1962); Roach (1961b); Manabe and Strickler (1964); Washington et al. (1967); and Eschelbach (1972) all have attempted to model the radiative characteristics of the atmosphere for solar energy. These models allow comparison between the observed results of the present study and those calculated from the models. They also allow us to estimate differences in absorption between a turbid and a purely gaseous atmosphere.

This study seeks answers to some of the questions concerning the magnitude and effect of solar energy absorption in the tropical atmosphere. Specifically these questions are: (a) what is the normal range of horizontal and vertical variation of absorption? (b) how does this energy input directly or indirectly affect tropical weather activity? and (c) what are the absorption, reflection and transmission characteristics of common cloud types and their effect on the atmosphere? The results presented here are a continuation of the preliminary work by Cox et al. (1968); Vonder Haar et al. (1969); Vonder Haar and Cox (1972) and Cox et al. (1973).

2.0 EXPERIMENTAL DATA

During the months of May, June and July, 1969, an aircraft radiation experiment program measured solar radiation absorption (.3 μm - 3 μm) in a tropical atmosphere. This experiment was part of the BOMEX carried out near the island of Barbados (59.5 W, 13.2 N).

The results presented in this paper were derived from pyranometer measurements of the total hemispheric irradiance measured by upward and downward facing instruments. Data used were from sensors mounted on five different aircraft:

(1) RFF Aircraft DC-6 40C DC-6 39C DC-4 82E	(2) NCAR Queen-Air	(3) NASA *Convair 990

*two modes of operation were used for determining solar convergence; see Fig. 1.

The single aircraft sounding method allowed versatility in the absorption measurements with respect to different atmospheric conditions. If an interesting cloud appeared or a clear area was sighted, the single aircraft could make a sounding in the area desired. With three or more aircraft stacked vertically this was not possible as the aircraft were flying specific flight tracks and levels designed to encounter a wide range of meteorological conditions. Fig. 2 shows typical pyranometric traces of downward radiation from all three RFF aircraft passing from a clear to a cloudy region.

The sensors used on each aircraft were Eppley precision spectral pyranometers. Specifications and calibration methods of these instruments may be found in Drummond and Hickey (1970). The modification of the sensors to aircraft use is described by Hanson et al. (1970). Since

TABLE I

DATE	TIME (LT)	ALTITUDE	WEATHER	AIRCRAFT	H↑ $\frac{WATTS}{m^2}$	H↑ $\frac{WATTS}{m^2}$
7/18/69	0745	5K'	Scat Cum	39C	430	49
"	"	"	"	40C	430	42
"	"	"	"	DC-4	465	49
"	"	"	"	Q-Air	472	49
7/23/69	0939	1.5K'	"	39C	840	63
"	"	"	"	40C	810	63
"	"	"	"	DC-4	860	77
*7/11/69	1435	1K'	"	Q-Air	730	35
"	"	"	"	CV990	745	28
7/13/69	1120	"	"	Q-Air	990	32
"	"	"	"	CV990	870	62

*Calibration made with Q-Air following the CV990 about five minutes behind at same altitude

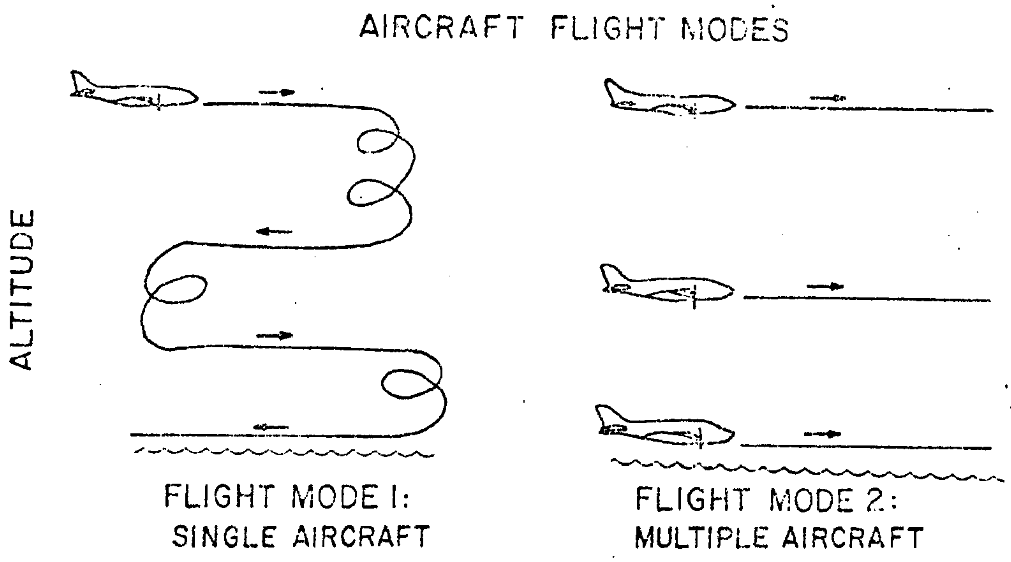

Figure 1 - Method of measuring the vertical convergence of solar radiation using single and multiple aircraft.

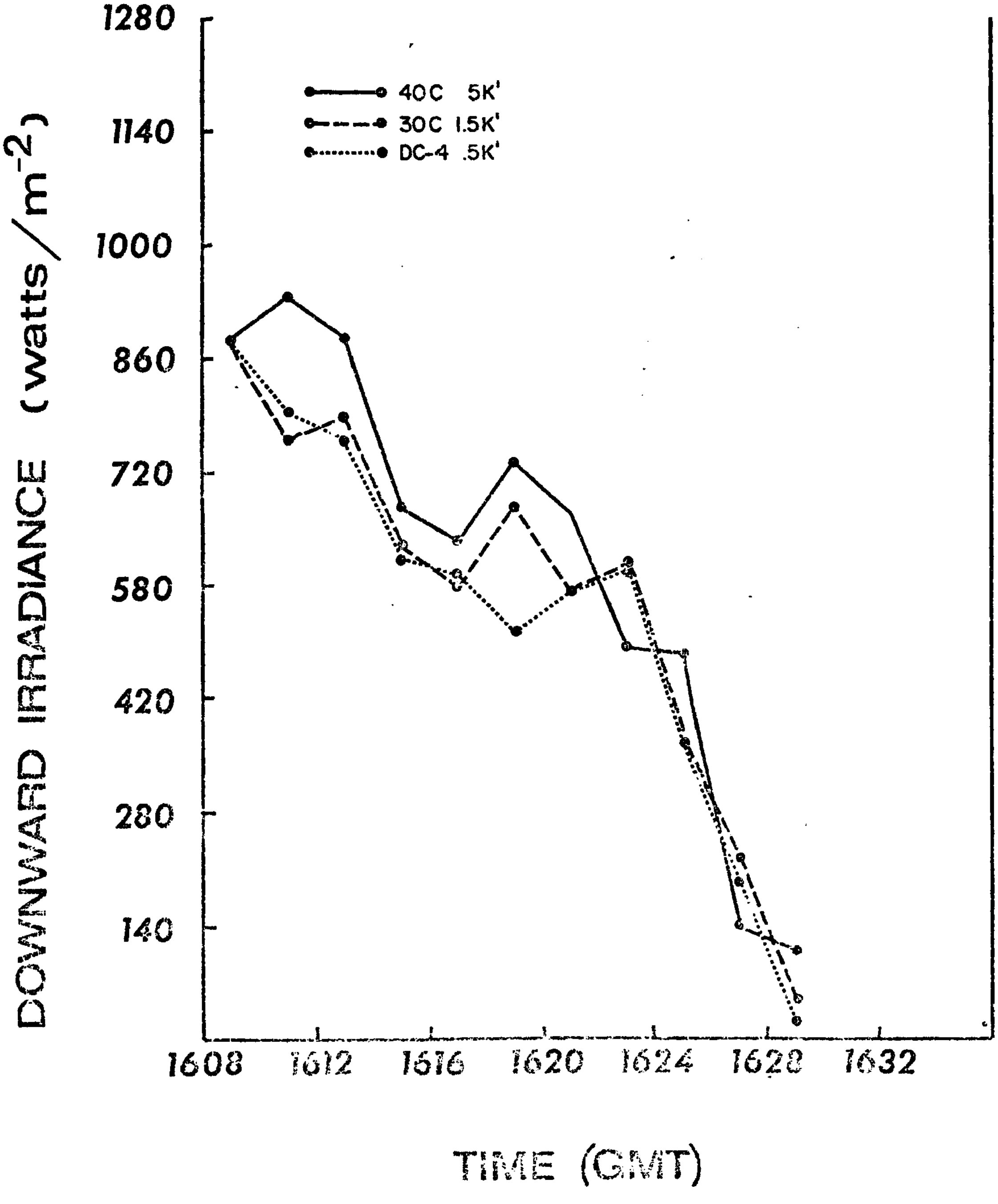

Figure 2 Downward solar irradiance as measured by the three RFF aircraft passing from a clear to a cloudy region.

our study sometimes used more than one aircraft to measure radiation convergence it was necessary to have repetitive intercomparisons between all the aircraft. This was done by having the aircraft fly in formation at the same altitude, then comparing the readouts of each instrument to the other (see Table 1). Corrections were made to the data when systematic differences were observed between the sensors by normalizing to a "standard". The data handling technique used in recording and calibrating the digitized radiometer output is given also in Hanson et al. (1970). Once the data were converted to energy units they were placed on magnetic tape. A statistical sampling of the data was then used to best represent the large number of values that were received during each "case"[1]; this gave approximately 150 values.

Time-lapse movies taken by the two RFF DC-6 aircraft were used in order to ascertain the extent of cloudiness of a region in which measurements were made. The two cameras were located on either side of the aircraft so that a relative estimate of the horizontal extent of the clouds could be determined. Combining the two aircraft films allowed location references to a cloud or clear area for each aircraft. In addition, observer's notes giving the meteorological conditions during the flights were used in the selection of clear or cloudy cases.

Satellite photographs taken by ATS-3 were also used for verification of measurement conditions. When a picture sequence was available over

[1] A "case" was determined by one sampling of clear air using either mode of aircraft operation or, one sampling of an individual cloud, again by either mode of operation.

the area of measurements the brightness field would indicate the spatial and temporal extent of a cloud or clear area where aircraft observations took place. Examples of the simultaneous satellite and aircraft measurements are provided by Vonder Haar and Cox (1972).

Standard meteorological and navigational information were automatically recorded on each aircraft. These data were also used in the course of the study of radiation data. More complete descriptions of the experimental conditions and data processing techniques are given in the references cited in the Introduction and by Reynolds (1973).

3.0 CASE STUDIES OF CLEAR CONDITIONS

3.1 Measurements

Cloudfree measurements were taken on 8 separate days and 14 "cases" were examined. Table two summarizes relevant information about these flights and the results obtained. The results were derived as follows:

↑		—— Upper (c)	Top T = Downward radiation watts-m^{-2}
z	3 [	2 —— Middle(b)	$\overline{x}$ = Average with time
		1 —— Lower (a)	Bot T = Upward radiation watts-m^{-2}

1) Absorbed radiation (watts m^{-2})

$$\text{Absorbed Radiation} = \{\, \overline{\text{Top T(c)}} - \overline{\text{Bot T(c)}} - \overline{\text{Top T(a)}} - \overline{\text{Bot T(a)}} \,\} \quad (1)$$

2) a) Absorbed radiation; percent of that incident on layer (%)

$$\%\ \text{Absorbed} = \frac{\text{Absorbed}}{\overline{\text{Top T(c)}}} \times 100 \quad (2)$$

b) Reflected radiation; percent of that incident on layer (%)

$$\%\ \text{Reflected} = \frac{\overline{\text{Bot T(c)}}}{\overline{\text{Top T(c)}}} \times 100 \quad (3)$$

c) Transmitted radiation; percent of that incident on layer (%)

$$\%\ \text{Transmitted} = 100\% - \{(a) + (b)\} \quad (4)$$

3) Heating rate (C/day)

$$\text{Heating rates} = -\, g/c_p \, \frac{\Delta H_{1,2,3}}{\Delta P_{1,2,3}} = \frac{\Delta T}{\Delta t} \quad (5)$$

4) Fractional Absorption

$$F_a = \frac{H{\downarrow}_{TOP} - (H{\uparrow} - H{\downarrow})}{S.C.} \qquad (6)$$

where:

S.C. = solar constant 1360 watts-m^{-2}

ζ_θ = zenith angle of sun at day, time and location of measurement

N = normalizing factor

5) Thickness change (meters day^{-1}) do to S.W. heating

Thickness change

$$\frac{\Delta z_1}{\Delta t} = \frac{Rd}{g} \frac{\Delta T}{\Delta t_1} \ell n \frac{P_a}{P_b} \qquad (7)$$

$$\frac{\Delta z_2}{\Delta t} = \frac{Rd}{g} \frac{\Delta T_2}{\Delta t} \ell n \frac{P_b}{P_c} \qquad (8)$$

$$\frac{\Delta z_3}{\Delta t} = \frac{Rd}{g} \frac{\Delta T_3}{\Delta t} \ell n \frac{P_b}{P_c} \qquad (9)$$

6) Change in stability (meters^{-1} day^{-1})

Equivalent Static Stability (Betts, 1973)

$$S = \frac{-10^{-3}}{\rho_a \dfrac{\overline{\theta}_{es(b)} + \overline{\theta}_{es(a)}}{2}} \quad \frac{\theta_{es(b)} - \theta_{es(a)}}{\overline{P}_b - \overline{P}_a} \quad \text{etc.} \quad \text{meters}^{-1} \qquad (10)$$

Change in Static Stability

$$\frac{\partial S^*}{\partial t} = \frac{-10^{-3}}{\rho_a \dfrac{\overline{\theta}_{es(b)} + \overline{\theta}_{es(a)}}{2}} \quad \frac{\Delta T}{\Delta t_1} - \frac{\Delta T}{\Delta t_2} \quad / \ \Delta P \qquad (11)$$

DATE 1969	TIME Z	PRESS MB	WATTS $H\downarrow m^{-2}$	WATTS $H\uparrow m^{-2}$	FRACTIONAL ABSORB.	(%) ABS.	(%) REF.	(%) TRANS.
June 26	1540Z	885 752 660	978 1009 1046	35 51 63	.28 .27 .25	4.0	6.0	90.0
June 26	1745Z	913 659	830 972	21 63	.29 .21	10.3	6.5	83.2
July 6	1600Z	899 703 597	940 1054 1112	28 73 84	.31 .26 .22	10.4	7.5	82.1
July 6	1515Z	913 802 599	936 959 1053	28 45 70	.31 .31 .25	7.2	6.6	86.2
July 2	1350Z	989 714	869 920	19 34	.25 .23	3.8	3.7	92.4
July 2	1505Z	989 714	1000 1063	20 38	.26 .22	4.2	3.6	92.2
July 18	1330Z	980 701 530	768 882 935	35 82 100	.34 .29 .26	11.0	11.0	78.0
July 18	1400Z	935 702 529	856 979 1041	35 87 107	.33 .28 .25	11.0	10.0	79.0
July 18	1500Z	933 701 529	1015 1091 1145	35 63 79	.28 .24 .21	7.5	6.9	85.6
July 18	1705Z	982 700 529	918 1020 1045	20 56 50	.25 .20 .18	9.3	4.7	86.0
July 23	1415Z	1006 856	874 918	35 35	.28 .25	4.8	3.8	91.4
July 23	1515Z	998 855	948 1023	28 42	.31 .27	6.0	4.0	90.0
July 11	1800Z	1000 700 312	781 946 1056	62 82 92	.32 .21 .13	25.3	7.9	66.7
July 20	1430Z	1000 780 650 450 270	882 1010 1093 1148 1222	72 92 123 113 113	.33 .25 .21 .16 .11	24.5	9.0	66.5

Table 2 - Results from radiation study in the cloudfree tropical troposphere.

$\Delta T/\Delta t$ $^{o}C/day$	MODEL HEATING RATE $^{o}C/day$	WEATHER	STATIC STAB. m^{-1}	$\Delta s/\Delta t$ m^{-1}-day^{-1}	OPTICAL DEPTH OF WATER VAPOR gm/cm^2	$\Delta z/\Delta t$ ft/day	$\Delta z/\Delta t$ m/day	AIRCRAFT
.90 2.4	2.1 1.8	HAZE	$-2.5x10^{-4}$	$+1.5x10^{-5}$	3.14	14.2 30.4	4.3 9.3	Q-Air
3.32	1.87	HAZE	$-2x10^{-4}$	N.A.	3.14	104.1	31.7	Q-Air
2.9 3.8	2.79 2.5	HAZE	$-2x10^{-4}$	$+6.5x10^{-6}$	3.54	69.5 59.5	21.2 18.1	Q-Air
.48 2.86	2.79 2.1	HAZE to 13 K'	$-2x10^{-4}$	$+1.6x10^{-5}$	3.54	5.4 80.5	1.8 24.5	Q-Air
1.09		HAZE	$-5x10^{-5}$	N.A.		34.1	10.4	RFF 82-39C
1.37		HAZE	$-4x10^{-4}$	N.A.		42.8	13.1	RFF 82-39C
2.02 1.72	1.97 1.5	HAZE ($90\mu g/m^3$)	$-9.1x10^{-5}$	$-3.1x10^{-6}$	2.42	65.1 46.2	19.8 14.1	RFF 82-39C-40C
2.57 2.03	2.03 1.53	HAZE ($160\mu g/m^3$)	$-1.1x10^{-4}$	$-5.9x10^{-6}$	2.42	70.9 55.4	21.6 16.9	RFF 82-39C-40C
1.73 1.88		HAZE ($40\mu g/m^3$)	$-1.0x10^{-4}$	$4.5x10^{-6}$		47.3 50.8	14.4 15.5	RFF 82-39C-40C
2.0 1.52		HAZE ($40\mu g/m^3$)	$-5.7x10^{-5}$	$-4.6x10^{-6}$		64.4 40.7	19.6 12.4	RFF 82-39C-40C
2.52	2.30	VERY HAZY	$-2.9x10^{-4}$	N.A.	2.53	38.2	11.6	RFF 82-39C-40C
3.60	2.49	HAZE	$-2.7x10^{-4}$	N.A.	2.53	53.2	16.2	RFF 82-39C-40C
4.04 2.37		HAZE BELOW 32 K'	$-7.34x10^{-5}$	$-1.1x10^{-5}$		139.4 184.2	42.5 56.1	CV990
4.14 3.37 2.75 3.44		HAZE BELOW 10.5 K'	$-4.8x10^{-5}$	$-5.6x10^{-6}$ $-4x10^{-6}$ $+4x10^{-6}$		98.88 59.07 97.09 108.7	30.1 18.0 29.6 51.4	CV990

Table 2 (cont.) Results from radiation study in the cloudfree tropical troposphere.

In order to compare instantaneous heating rates for different times and days it was necessary to first understand the physical factors that control solar radiation convergence (see figure 3). They are: 1) the insolation at the top of the atmosphere at the day and time of the measurements; 2) the geometry of the absorption, which would consider the path length of the radiation in the atmosphere (including the effect of multiple scattering; and 3) the density of the absorbing medium which would relate to the degree of turbidity (aerosol concentrations) for a given case (also includes multiple scattering effects). In order to separate some of these effects the instantaneous heating rates were normalized to the incoming solar radiation normal to the top of the atmosphere for the specific day, time and location of the observations as follows:

$$H_{\text{Top of Atmosphere}} = \text{S.C. x} \cos \zeta_\Theta = N \qquad (12)$$

$$\frac{\Delta T}{\Delta t_n} = - g/c_p \frac{\Delta H/N}{\Delta P} \qquad (13)$$

Hence the normalized heating rates vary only as a function of factors (2) and (3) stated above.

Separate measurements of the scattering which took place during the period of aircraft measurements were not available nor were there good estimates of the turbidity for all cases. Measurements of aerosol concentrations for some layers were made by Prospero and Carlson (1972) and when thermodynamic soundings were available near the radiation measurements the water vapor optical mass was calculated. These supporting data will be considered later for specific case studies.

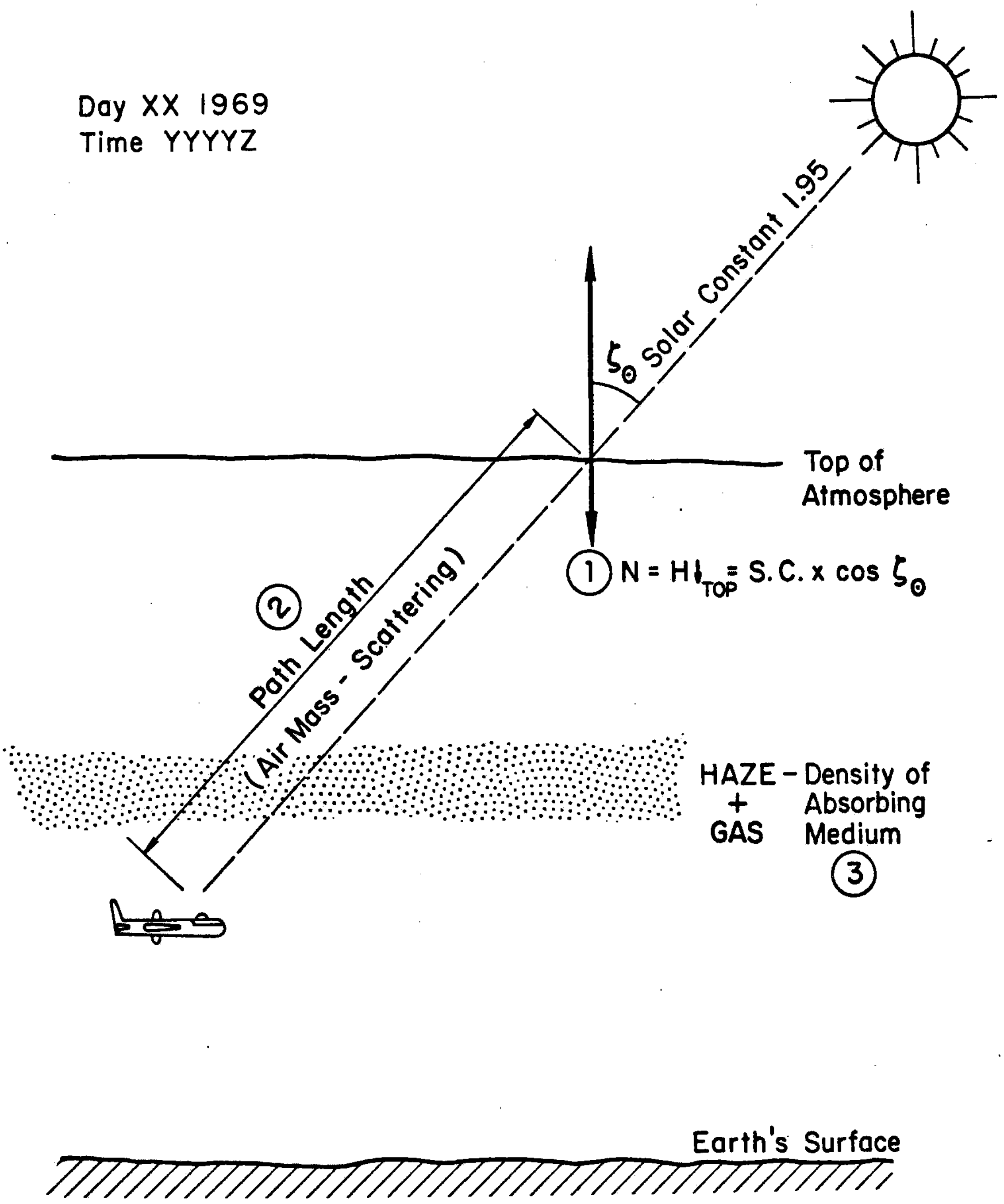

Figure 3 - Geometry of solar insolation and factors causing attenuation of the incident beam.

3.2 Results

Fourteen clear (cloudfree) cases were studied and all measurements were carried out for layers of .5 km depths or greater due to the sensitivity of the pyranometers used. It was found that when layers of less than .5 km were used the attenuation of radiation was small (near the noise limit of the system) and the effects of scattering large such that erroneous absorption values were obtained. This factor was the only limitation to the measurements made in clear air. The results show that in the lower half of the tropical troposphere (1000 - 500 mb) approximately 9% of the incoming solar radiation incident on this layer is absorbed, 8% is reflected and 73% is transmitted.

Figure 4 shows vertical profiles of heating rates composed from all BOMEX data for the clear tropical troposphere. In the lower 200 mb a large scatter was found in the heating rates derived from measurements of the amount of absorbed solar radiation. There seems to be a slight maximum in the heating rate profile around 700 mb but the significance as a composite representation is dubious with only three data points available near that pressure level. Above the 600 mb layer the heating rate profile seems to be fairly constant up to 400 mb which was the midpoint of the uppermost layer of our measurements.

The large scattering of heating rates in the lower 200 mb of the tropical atmosphere may be partly explained by considering the measurements made on 18 July from the RFF aircraft. According to Carlson and Prospero (1971) large Saharan air mass outbreaks made up of small particles of sand may be carried in the wind flow in the 600 - 800 mb.

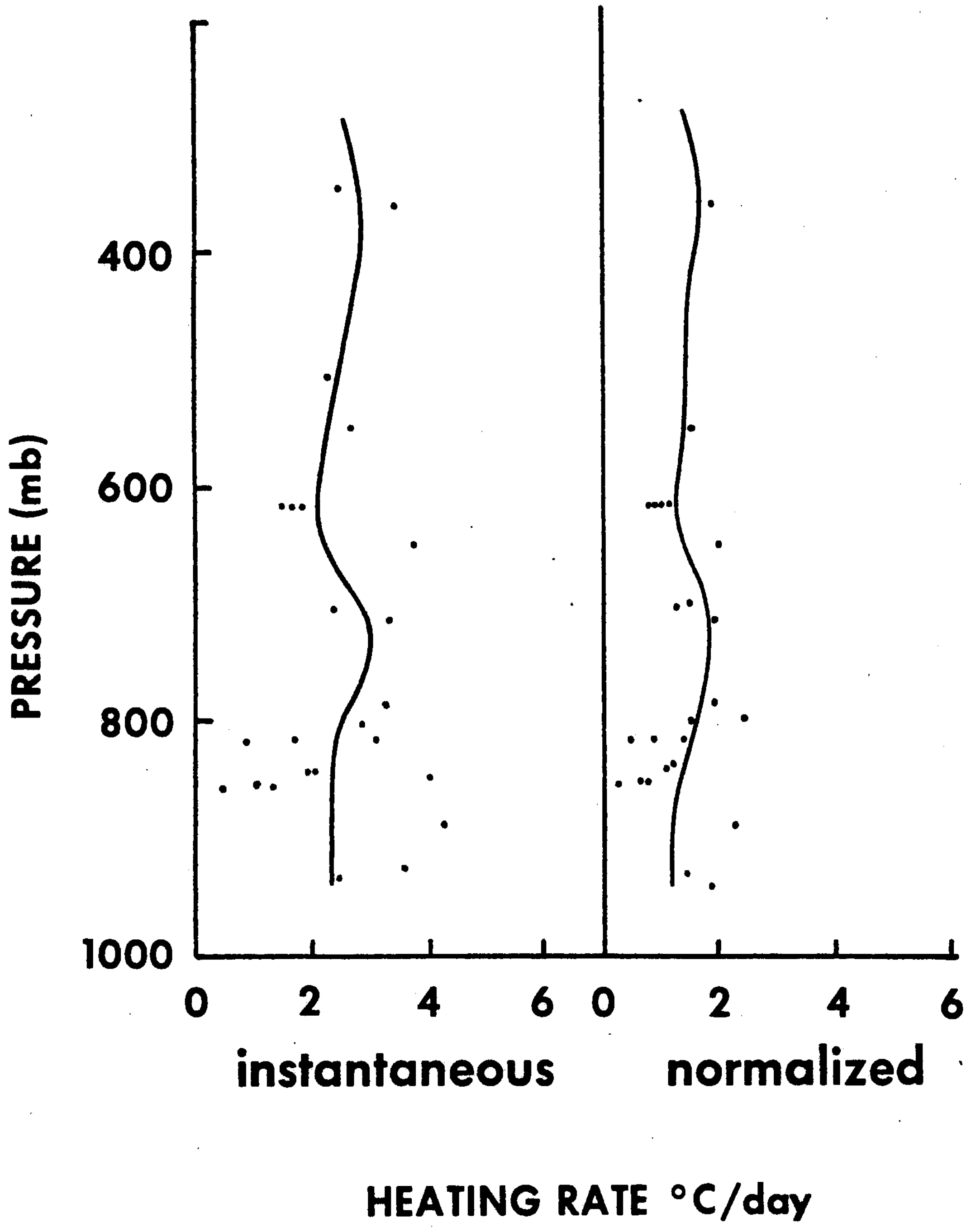

Figure 4 - Instantaneous and normalized heating rate profiles using a hand-drawn fit to the data points. A slight maximum in the heating rates may exist between 800-600 mb. The normalized data exclude the effect of varying insolation. Profiles are composed from all cloudfree measurements obtained during BOMEX.

level to the BOMEX area. Their measurements indicate that high concentrations (>150 $\mu g/m^3$) of suspended matter (minerals) may be found in the air over the BOMEX array and make up a substantial part of the haze found in the tropics. Prospero and Carlson (1972) carried out measurements during the BOMEX to try and estimate the magnitude of this turbidity. Particulate concentration, measured by high volume air samplers, was determined during the radiation convergence flights of the RFF DC-6 aircraft on 18 July. On this date the RFF 40C flew at the 700 mb level as the middle aircraft in the vertically stacked mode for radiation convergence measurements. According to Prospero this would put 40C in the Saharan air layer where Prospero found the average mineral aerosol concentrations to be 61 $\mu g/m^3$ for the three month period of the BOMEX.

Particulate concentrations for the specific clear cases of 18 July are shown in Table 2. At 1400Z, one of our radiation cases, the 160 $\mu g/m^3$ concentration was the highest measured by Prospero during the entire BOMEX experiment; both the films and observer notes indicate turbid conditions (see Fig. 5). Figure 6 presents a graph of fractional absorption for the four cases of July 18 showing the increase for the turbid conditions. For comparison we also show a profile of the fractional absorption for a clear, gas-only tropical atmosphere as calculated using the method of Manabe-Strickler (1964; see section 3.3) and a profile for the mean fractional absorption measured for all clear cases of our study (data base the same as Figure 4). These values show quantitatively the additional solar radiation attenuation due to the suspension of aerosols in a layer. Thus the presence or absence of aerosols in the lower layer may greatly affect the amount of absorption taking place, and in the

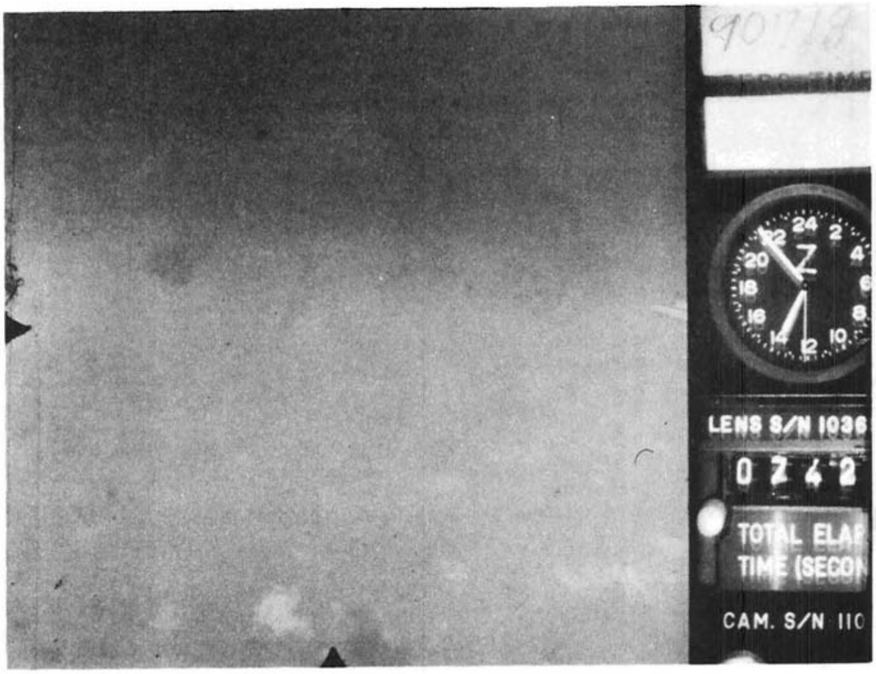

Figure 5 - Single frame from a time lapse movie film taken onboard 40c at 1358Z 18 July showing extreme haze and dust.

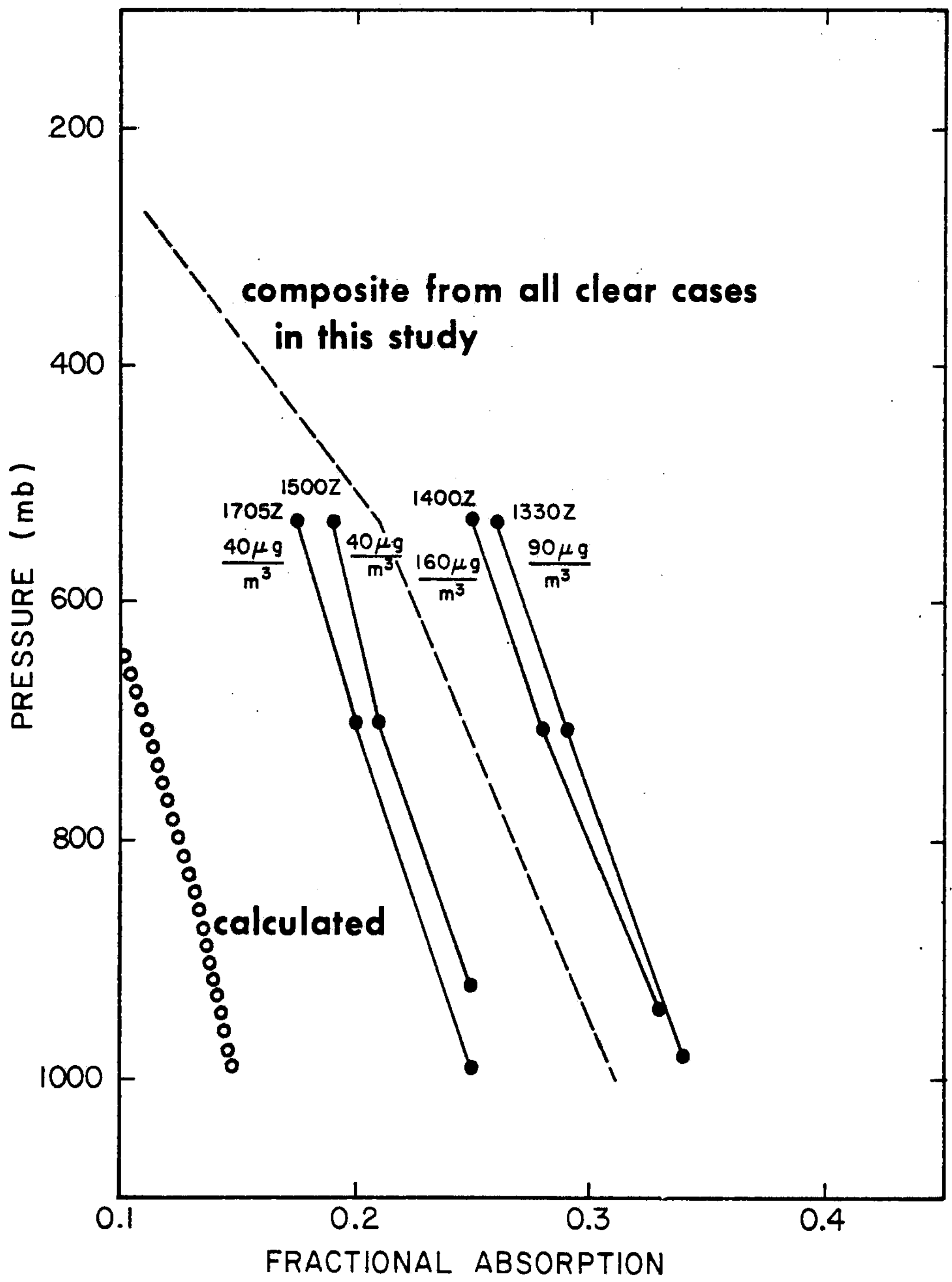

Figure 6 - Fractional absorption in the tropical atmosphere showing increased absorption for the more turbid cases. Comparison to theoretical calculations showing an apparent increase in absorption over that accounted for by the gaseous constituents of the atmosphere.

absence of large water vapor content variations, most of the scatter in the lowest layer cloudfree measurements noted in Figure 4 is probably indicative of aerosol variation.

Columns 13, 14 and 16, 17 of Table two give values for the atmospheric stability and change in stability as well as thickness changes which would occur if the radiation heating rates for the daylight hours (see summary) went directly to sensible heat for the atmosphere. These parameters are used as an illustration of the possible effect the vertical solar heating may have on the atmosphere. The actual stability is given in column 17 so that a comparison of it to the inferred stability change can be made.

3.3 Use of Theoretical Model for Comparison with Measurements

A method to estimate the effects of aerosol concentrations on solar radiation convergence is to compare the observed heating rates to results from theoretical models which handle the gaseous constituents in the atmosphere but not foreign materials. The model used for comparison was one developed by Manabe and Strickler (1964) which uses an atmospheric sounding of temperature and moisture (a mean CO_2 mixing ratio was also used). Thus, it was necessary to have a radiosonde launch concurrent with the time of our radiation measurements and consequently only those measurements taken within 150 miles of an upper-air station (ships or Island of Barbados) were able to be compared with the theoretical values.

Figures 7 and 8 show the profiles of the measured upward, downward and net solar radiation and the downward radiation calculated by the model for all cases available (8 of the 14). In each case the model overestimates

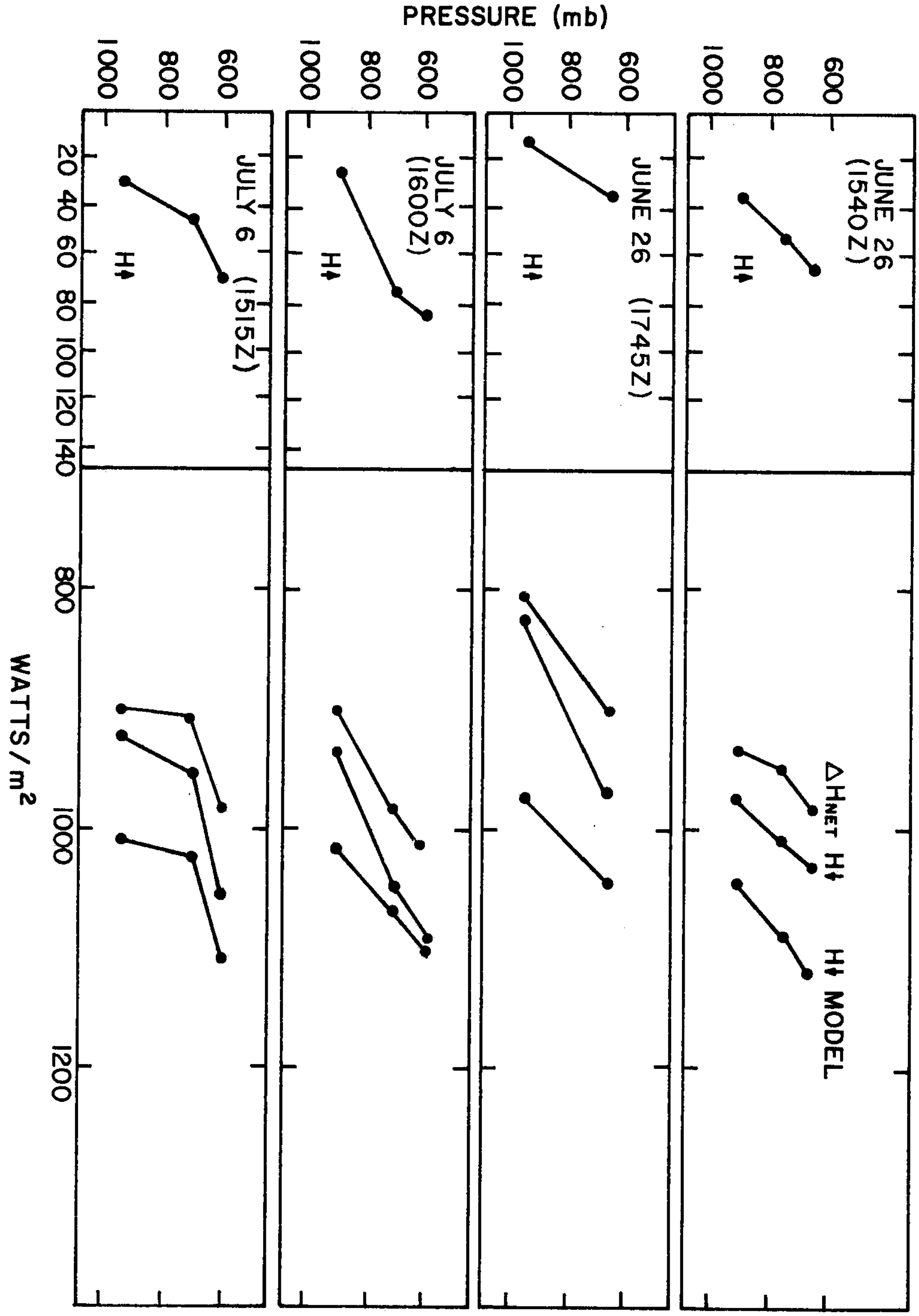

Figure 7 - Vertical profile of the upward, downward and net radiation and vertical profile of the downward radiation as calculated by theory showing theory consistently calculates a larger downward flux than observed.

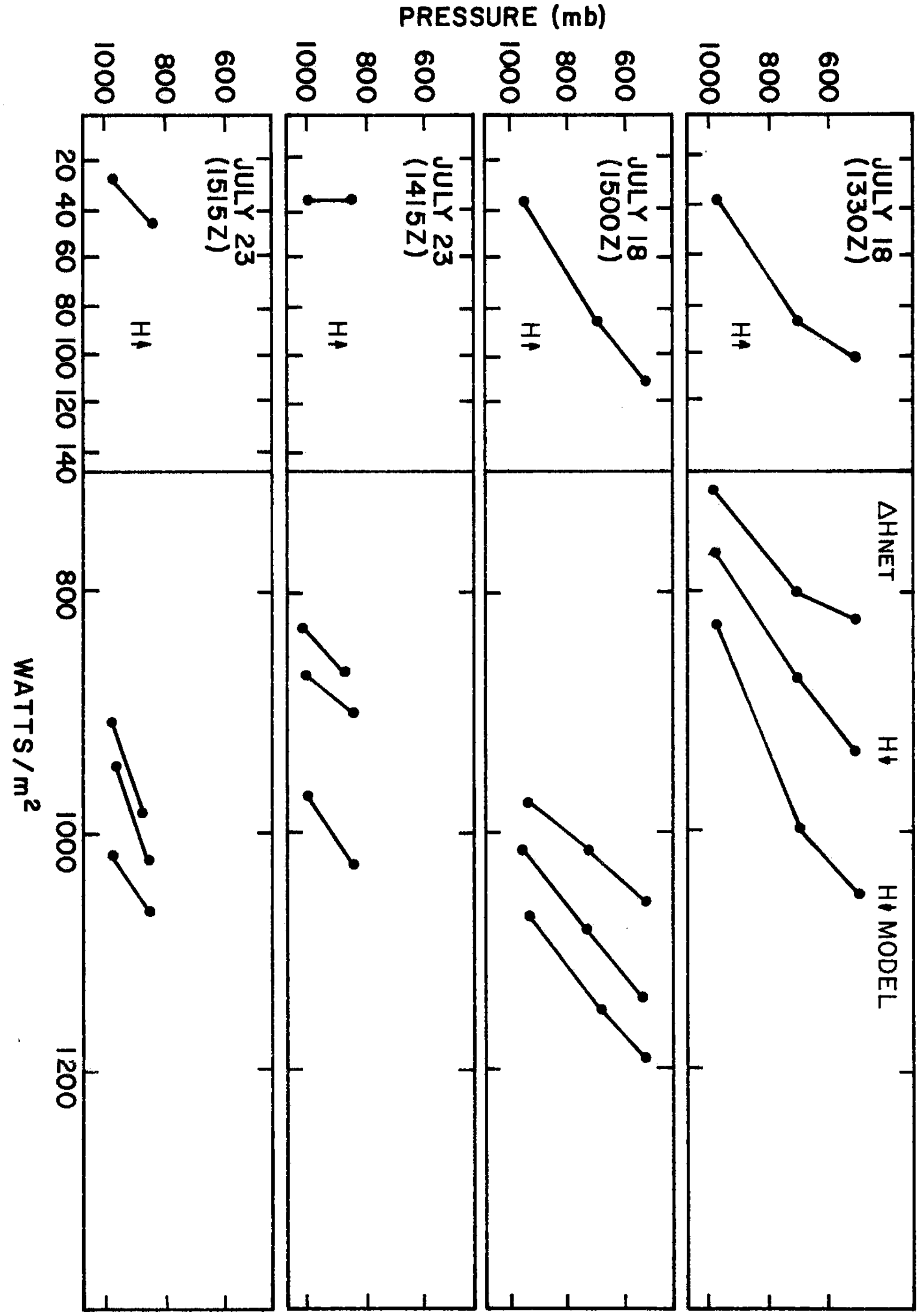

Figure 8 - Vertical profile of the upward, downward and net radiation and vertical profile of the downward radiation as calculated by theory showing theory consistently calculates a larger downward flux than observed.

the downward radiation indicating that either more absorption is taking place or that more solar radiation is being reflected to space than accounted for by the model.

The use of this simple theoretical model together with measurements of radiation has shown that when aerosols and a large water vapor concentrations are present in the tropical atmosphere they amy act to absorb the solar radiation along with the normal absorption by the gaseous constituents in the atmosphere. This absorption seems to occur more above the 800 mb level and can be correlated to the Saharan air layer between 800-600 mb found by Prospero and Carlson. Kondratyev et al. (1972a), in a study similar to this one, found that the atmospheric flux convergence due to the absorption by aerosols and the atmospheric gaseous components may be of equal magnitude.

3.4 Role of solar radiation in the energy budget of the cloudfree tropics

The authors attempted to assess the role of solar radiation in the clear tropical atmosphere by following the earlier work of Yanai (Yanai et al., 1972) who tried to determine the interaction between large-scale sensible, latent and radiational heating or cooling and tropical waves or cloud clusters. Yanai considered an ensemble of clouds which are embedded in a large scale tropical weather system. He then used budget equations for a layer of the atmosphere based on the equations of mass continuity, first law of thermodynamics, and moisture continuity, all averaged over a horizontal area large enough to contain the cloud ensemble but small enough to be a fraction of a large scale system.

The budget equations are:

$$^{*}\quad Q_1 = Q_R + L(c-e) - \frac{\partial}{\partial p}\ \overline{s'w'} \tag{14}$$

$$^{*}\quad Q_2 = L(c-e) + L\frac{\partial}{\partial p}\ \overline{q'w'} \tag{15}$$

where:

$\overline{w}$ = average vertical p-velocity

Q_R = heating rate due to radiation

c = rate of condensation per unit mass of air

e = rate of re-evaporation of cloud droplets

$\overline{(\ )}$ = horizontal average

s = c_p + gz (dry static energy)

$(\ )'$= small scale eddies

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*The eddy terms of (14) and (15) refer to the actions of mesoscale convective clouds in the tropical environment.

Eq. (14) shows that the apparent heating of the large scale motion system (Q_1) consists of heating due to radiation, the release of latent heat by net condensation and vertical convergence of the vertical eddy transport of sensible heat. Eq. (15) is the moisture continuity expressed in units of heating rate, and is a measure of the apparent moisture sink in a layer (Q_2) which is due to the net condensation and vertical divergence of the vertical eddy transport of moisture. Yanai derived from the above two equations the following:

$$Q_1 - Q_2 - Q_R = -\frac{\partial}{\partial p}\ \overline{h'w'} \tag{16}$$

where $\overline{h'w'}$ equals a measure of the vertical eddy transport of total heat or moist static energy; a measure of activity of cumulus convection.

This equation should hold for any layer in the atmosphere thus allowing the energy requirements of each layer to be inferred. For a layer high in the troposphere (i.e., 500-200 mb) the latent heat terms would be quite small due to the lack of moisture. Thus the balancing terms would be the sensible and radiation terms which would be balanced by induced vertical motions (Gray, 1972). When the atmosphere is taken as a whole, all terms become important to the energy balance.

Yanai shows that if equations (14) and (15) are integrated from sea surface to cloud top that estimates of Q_1 , Q_2 and Q_R can be checked by surface measurements of precipitation, sensible heat supply from the ocean and rate of evaporation from the ocean. For a case in the Pacific that he studied:

$$1/g \int_{P_T}^{P_0} (Q_1 - Q_R)\, dP = LP_0 = \text{precipitation } (1086 \text{ cal}\cdot\text{cm}^{-2}\cdot\text{day}^{-1}) \tag{17}$$

$$1/g \int_{P_T}^{P_0} Q_2\, dp = L(P_0 - E_0) = \text{precipitation-evaporation } (697 \text{ cal}\cdot\text{cm}^{-2}\cdot\text{day}^{-1}) \tag{18}$$

where:

L = latent heat of vaporization

P_0 = precipitation

E_0 = evaporation

Figure 9 shows a vertical profile of Q_1 and Q_2 as given by Yanai from data from the Marshall Island Experiments of 1956. The vertical profile of Q_R is also given as derived by Dopplick (1970) from climatological values, assuming two-tenths cloud cover. The graph also shows the vertical profiles of infrared cooling Q_L[1] along with the observed mean solar heating rates for cloud areas[2] Q_S during BOMEX. For the vertical eddy transport of total heat equal $\dot{Q}$ Yanai estimates it would take 1.8 cm/day of precipitation rate was 1 cm/day. This discrepancy may be resolved, as Yanai points out, by the fact that the radiational heating as Dopplick calculated may be too low. If we assume for the daylight period[3] that the solar heating amounts to 2.5 C and the longwave cooling is .75 C (one-half an estimate for a day in the tropics; Cox, 1972), then following Yanai's approach

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[1]This curve does not take into account the fact that water vapor pressure broadening occurs in the atmospheric window, 8-12 pm, such that the infrared cooling in the tropical troposphere increases by as much as 30% (Cox, 1973). This would somewhat reduce the values in the Q_R curves, especially in the lower 200 mb.

[2]The cloudfree heating rates are used as a first approximation to Yanai's budget box since it is felt the clear areas make up approximately 80% of the total area. Future work will consider the results from clouds to better approximate the true conditions in the area considered. The instantaneous mean values shown in Figure 4 have been divided by two to account for the relative amount of daytime and nighttime, since all profiles in Fig 9 apply to the average daily case.

[3]The use of the mean instantaneous heating rate as a value for the daylight solar heating of the atmosphere follows from a simple argument that the increased path length of the solar radiation through the atmosphere with decreasing solar elevation maintains the heating rate at its instantaneous value to the first approximation. Kondratyev (1972) has found during the CAENEX (Complex Atmospheric Energetics Experiment) that the maximum solar heating in the troposphere may not occur at local noon but rather in the forenoon or afternoon when dust loading is highest

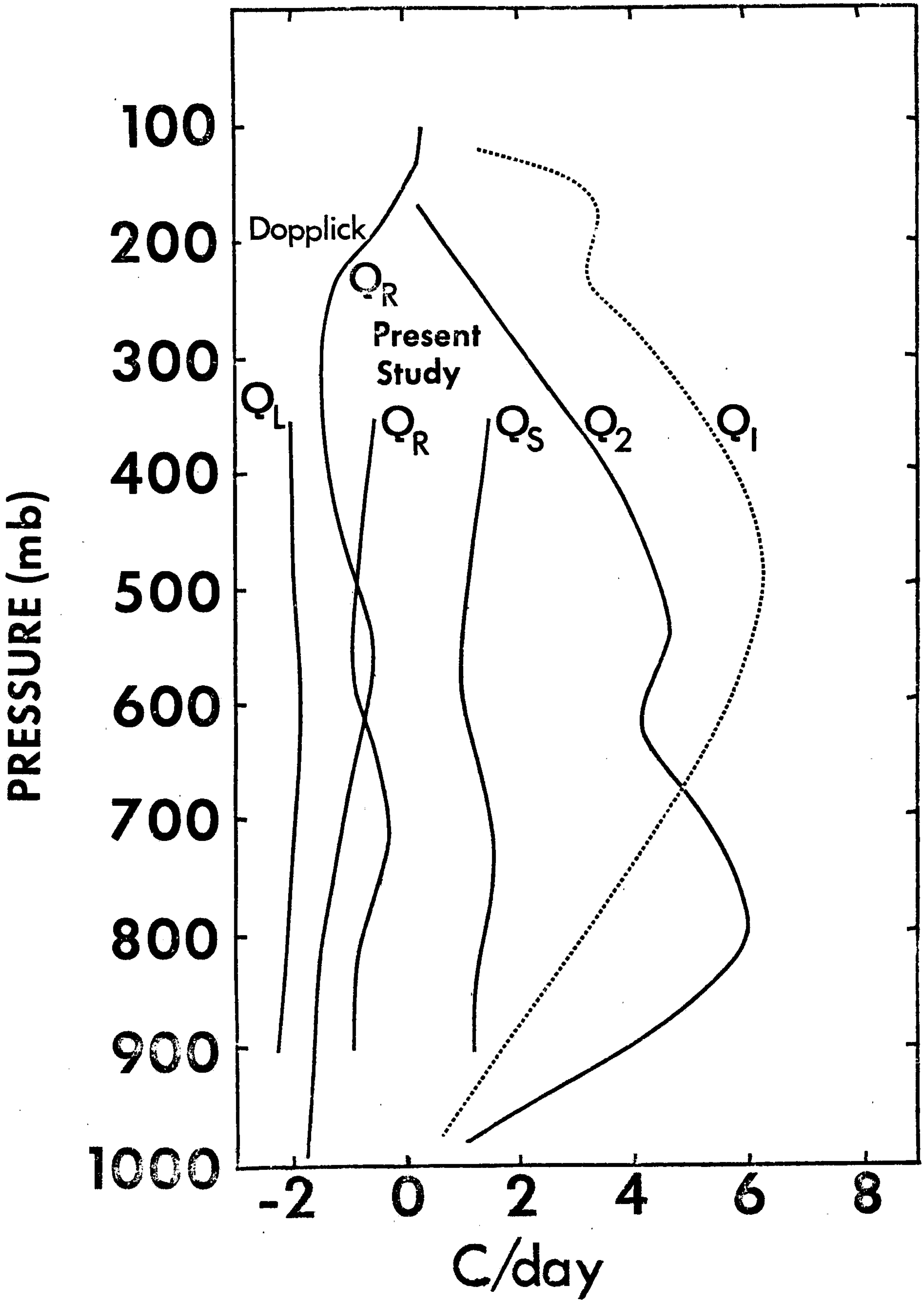

Figure 9 - Vertical profiles of total heating (Q_1), moisture continuity (Q_2), solar heating as found in this study (Q_S), net heating using Q_S (Q_R Present Study), net heating as found by Dopplick (Q_R Dopplick), and longwave cooling of the atmosphere (Q_L) found by Dopplick.

the precipitation would amount to .25 cm for the 12 hour daytime period. During the nighttime hours when only infrared cooling is taking place, .93 cm of precipitation would be needed to balance the sensible and latent heating terms, and the daily sum would be 1.18 cm which is .18 cm higher than the amount observed at the surface during the period in which Yanai carried out his study. Yanai points out that the precipitation measurements made were probably too low for this large an area and also that the measurements of sensible and latent heat fluxes may be off by a factor of 1.5 to 2.0. The result inferred from the present argument is .64 cm/day closer to the observations than that found by Yanai using Dopplick's heating rates. Thus, this says that there may be a close relationship between the radiation balance of large areas in the tropics and the precipitation. For the very long time averages this has been acknowledged for many years (e.g., Riehl and Malkus, 1958). However, could the relation also exist on such a short time scale as the diurnal? Perhaps it does when larger scale migratory disturbances are not considered since it has been noted that the precipitation maximum does occur in the tropics during the nighttime hours (Holland, 1970; Atkinson, 1971) in direct correlation to what would be required from these energy balance considerations. In order to test this speculation more radiation measurements along with sensible and lantent heat flux measurements are needed to estimate the magnitude of each term and then correlate this to the precipitation amounts. This work could easily be carried out in the GATE (GARP Atlantic Tropical Experiment) program planned for 1974.

4.0 RADIATION CASE STUDIES OF CLOUD CONDITIONS

4.1 Background

Another aspect of the solar radiation studies carried out during the BOMEX was the study of the radiation characteristics of certain cloud types in the tropics. Clouds undoubtedly dominate the radiation budget of the tropical atmosphere where they are present in large amounts. The questions that are posed in this study are: (1) what is the magnitude of solar energy absorption by different cloud types; (2) does the absorption act in a way to inhibit or encourage cloud growth; and (3) what is the effect of the cloud radiation characteristics on layers of surfaces underneath or above, the cloud layer itself.

Measurements applied to answer the above points were made using the five aircraft previously mentioned. The number of cloud cases available for this work was limited due to the aircraft operations themselves. In most cases the clouds used for this study were not directly sought out for measurement by the aircraft but were encountered along a preplanned flight track. Thus there were only eight acceptable cloud cases available for this work.

4.2 Results

Table three shows the results of the eight different cloud measurements made during this experiment. The only cloud types measured were stratocumulus, cumulonimbus or congestus, and cirrus. Three measurements were made of strato-cumulus clouds, three of cumulus clouds and two of cirrus. It must be mentioned here that only in a few cases were the exact depth and width of the cloud measured and the dimensions of

DATE	TIME (GMT)	LAT.- LONG.	PRESS. (MB)	WATTS Wm^{-2}	WATTS Wm^{-2}	FRACT. ABS.	% ABS.	% REF.	% TRANS.	$\Delta T/\Delta t$ $^oC/day$	$\Delta T/\Delta t_n$ $^oC/day$	NORMAL FACTOR $ly\text{-}min^{-1}$	$\Delta T/\Delta t_{L.H.}$ $^oC/day$
June 9	1527 to 1533Z	13.7 N 59.9 W	815 698	511 986	54 414		12	42	46	8.3	4.37	1.89	5.25
July 2	1657 to 1701Z	15.2 56.3	989 528	312 1123	7.7 410		36	37	27	7.5	3.85	1.94	2.75
July 18	1429 to 1432Z	16.07 52.2	990 700 529	478 1063 1116	24 396 412	.64 .48 .45	22	37	41	6.2 1.8	3.31 .98	1.87	2.6
July 14	1530 to 1536Z	15.1 53.4	977 716 295	186 600 1217	7.0 121 530	.89 .66 .49	42	44	14	9.7 4.2	5.01 2.2	1.94	3.5 2.6
July 14	1451 to 1510Z	14.7 54.7	724 295	287 1217	17 457		52	38	10			1.93	5.7
July 23	1810 to 1815Z	8.8 51.2	966 661 235	21 40 785	0.0 7 521	.75 .74 .55	31	66	3	.35 4.6	.24 3.15	1.45	.13 3.01
July 18	1707 to 1716Z	11.32 53.32	200 355	1075 513	6360 230		14	59	27	8.41	4.75	1.77	7.76
July 13	1652 to 1707Z	7.02 53.0	300 640 1010	1139 625 496	534 62 41	.46 .49 .57	13	47	40	1.05 2.46	.61 1.42	1.74	.67

DATE	TIME (GMT)	STATIC STAB. m^{-1}	$\Delta S/\Delta t$ $m^{-1}\text{-}day^{-1}$	$\Delta S/\Delta t_{L.H.}$ $m^{-1}\text{-}day^{-1}$	$\Delta z/\Delta t$ ft D^{-1}	$\Delta z/\Delta t$ m D^{-1}	$\Delta z/\Delta t$ ft D^{-1}	$\Delta z/\Delta t_{L.H.}$ m D^{-1}	CLOUD TYPE	AIRCRAFT	COMMENTS	$^oC/day/mb$ $\Delta T/\Delta t/\Delta P$	$^oC/day/mb$ $\Delta T/\Delta t/\Delta P_n$
June 9	1527 to 1533Z	$-1x10^{-4}$	N.A.	N.A.	122.9	37.5	78.0	24.0	St. Cu.	Q-Air		.40	.21
July 2	1657 to 1701Z	$-6x10^{-5}$	N.A.	N.A.	450.4	137.3	165.7	50.5	St. Cu.	RFF		.20	.10
July 18	1429 to 1432Z	$-7.7x10^{-5}$	$-4.2x10^{-5}$	$-1.0x10^{-5}$	205.9 49.2	62.8 15.0	67.2	20.5	St. Cu.	RFF		.64	.34
July 14	1530 to 1536Z	$1.2x10^{-5}$	$-1.2x10^{-5}$	$-3.25x10^{-5}$	289.8 365.0	88.3 111.0	104.2 231.0	31.8 70.4	Td	RFF CV990	below cloud in cloud	.037 .010	.02 .005
July 14	1451 to 1510Z	$2.8x10^{-5}$	N.A.	N.A.	830	252	491	149	Td	RFF CV990	in cloud	.022	.01
July 23	1810 to 1815Z	$5.9x10^{-5}$	$2.4x10^{-5}$	$1.7x10^{-5}$	12.6 454	3.8 138	4.7 295.0	1.4 90.0	Cb	RFF CV990	in cloud rain in cirrus shield	.001 .010	.0007 .007
July 18	1707 to 1716Z	$2.3x10^{-4}$	N.A.	N.A.	463.5	141.3	427.8	130.4	Ci	CV990		.51	.035
July 13	1652 to 1707Z	$-9x10^{-6}$	$-9x10^{-5}$	$-1.7x10^{-5}$	76.59 101.9	23.3 31.1	48.8	14.9	Ci	CV990		.084	.015

Table III Results from radiation study of certain cloud types in the tropics.

most clouds cited in this study can only be estimated. This was done by means of a visual estimation of the cloud's depth using films and slides taken on measurement days. These estimates are given in Table 3.

When aircraft making cloud measurements flew substantially above or below cloud top or base, a correction was applied to the absorption percentage found due to the fact that not only was cloud absorption taking place but clear air absorption was taking place as well. Thus the observed absorption percentage was decreased by an amount determined from measurements made in clear air under similar conditions and similar depths of atmosphere. Then the recomputed absorption percentages, cloud minus clear air, were used to recalculate heating rates that should correspond more closely to the heating in the clouds themselves. Taking these values and dividing them by the depth of the cloud in millibars, an estimate of the heating per millibar of the cloud can be found. These values are also shown in Table 3 (cols. 24 and 25).

It is understood that the radiative heating which has been discussed may be manifested into many different processes in the cloud layer in which the absorption takes place. These processes include induced vertical motions and horizontal wind, sensible heating, evaporation of cloud water and transport out of the absorbed solar energy. A simple technique developed by Knollenbert (1972) may allow us to separate one of the above processes from the rest due to the face we are dealing with a cloud. The following equation (19) allows an estimate of the amount of evaporation of cloud water induced by the abserved solar energy thus giving another heating rate which would supposedly be the energy available for the other processes to take place.

$$\frac{\Delta T}{\Delta t} = \frac{g\,\overline{\rho}_a}{\overline{\rho}_a c_p + \overline{\rho}_s L^2/(R_w \overline{T}^2) - L\overline{\rho}_s/\overline{T}} \qquad (19)$$

[function of T_{cloud} only]

where:

g	=	$980\ cm/sec^2$	L	=	$2.5 \times 10^{10}\ ergs/g^oK$
R_d	=	$2.87 \times 10^6\ ergs/g^oK$	ρ_s	=	saturation vapor density f(T)
R_w	=	$4.6 \times 10^6\ ergs/g^oK$			
c_p	=	$1 \times 10^7\ ergs/g^oK$	ρ_s	=	$1.2 \times 10^{-3}\ gm/cm^3$

These corrected heating rates are again shown in Table 3 (col. 14).

Table four summarizes the results of the cloud study made here and the results of two other studies. The case for the strato-cumulus clouds serves as an example of the wide range of natural variability in the radiative characteristics of similar cloud types (absorption 12-36%). Fig. 10 simply represents the vertical profile of solar radiance made of three different cloud types during the BOMEX. These type of profiles along with past and future measurements should help obtain usable limits to the radiative properties of clouds.

4.3 Atmospheric Effects due to Radiative Processes in Clouds

Since clouds have been shown to absorb a significant amount of solar radiation, this absorption may affect the cloud's growth as well as the cloud's environment. The fact that some of the absorbed radiation goes into evaporating water has been discussed and is shown to play a substantial role for clouds low in the atmosphere. Gille and Krishnamurti (1972) have shown that when radiative cooling by longwave radiation is placed into a model of a tropical disturbance the energetics are changed in such a way as to enhance the conditional instability of these disturbances. This study will try to show that solar radiative warming may affect the stability of cloudy regions in a similar manner.

TABLE IV Summary of the radiative properties of certain cloud types using the results from this study and two others.

TYPE	REFLECTED (%)	TRANSMITTED (%)	ABSORBED (%)
Cu (large, thick)	38-44	10-14	36-42 [25]*
Cb	66	3	31
St-Cu	37-42 [47]* (60)*	27-46	12-36
Ci-Cs	47-59 [20]* (20)*	27-40 [80-40]*	13-14 [10-30]*

*( ) London (1957) mean

*[ ] Drummond and Hickey (1971) mean

NOTE: All solar elevation angles greater than 60°

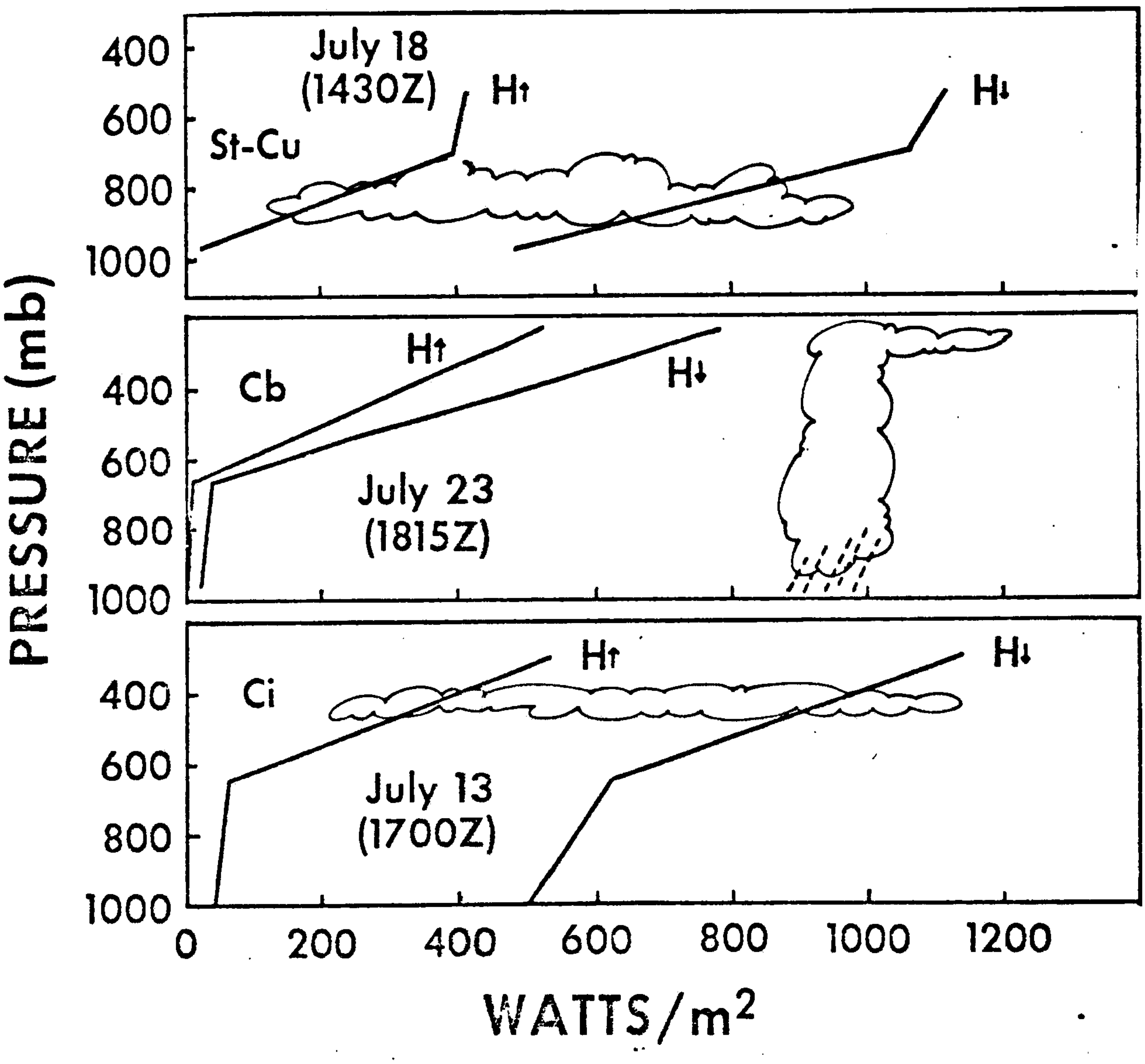

Figure 10 - Upward and downward components of solar radiation through Strato-cumulus (St-Cu), a cumulonimbus (Cb) and through cirrus (Ci).

The strato-cumulus cloud measurements of 18 July provided the only case of this cloud type where sufficient data were available to calculate the change in stability due to the heating rate profile. The data shows that with the cloud located in the lower part of the atmosphere the differential vertical heating would act to destabilize the whole layer (990-529 mb) over a day.

When the modified heating rate (corrected for latent heat affects) is used it again shows a destabilization occurring but to a lesser extent. Thus, it may be concluded that the solar radiation acts in such a way as to enhance the growth of this cloud. However, what is probably occurring is that in the upper portion of the cloud more absorption takes place than in the lower half. This would increase the evaporation of cloud water and cool the upper portion of the cloud more than below which would enhance the destabilization. Since this is a stratus type cloud we can assume it is capped by a more stable or drier layer such that with increased growth of the cloud more mixing with this drier air would occur and thus erode the cloud.

A related question posed here concerns the effect at the ocean air interface when the solar insolation is drastically reduced by the large cumulus clouds (i.e., July 23 Cb case). Hanson (1971) has shown that when a large cloud system exists over an area for an entire day the insolation at the ocean surface may be reduced by as much as 99%! This obviously would affect the sensible heating of the ocean surface as well as the air above. Also, evaporation from the ocean surface may be reduced. The short-term affects this may have on the mesoscale dynamics of the area may be such as to try and dissipate the cloud system and bring the energy balance back into phase. This is am important question that must be looked into further since such a large amount of energy is being considered (1000 $cal \cdot cm^{-2} \cdot day^{-1}$).

In discussing cirrus clouds we may assume here that the energy absorbed by the cirrus may be as large as the energy released by forming the cirrus. A simple calculation for a cirrus cloud 100 km^2 in area and 1 km deep shows the absorbed radiation over a day (10^{12} cal) is of the same magnitude as the energy required to maintain the cirrus cloud over a day. Thus the effect the solar heating has on the cloud and its local environment may be quite important, especially if the cloud was to maintain itself for an extended period, say one week. The cirrus cloud measured on 18 July probably shows the most realistic relationship between cirrus and the incident solar radiation. Here we see a rather large absorption taking place shown by the 8.41^{o}C/day heating rate. With such a large depletion of the incident radiation, only 27% transmitted, little is left to be absorbed below the cloud. Thus, very little heating below the cloud would result in a large stabilizing effect by the cloud. Consequently the cirrus not only warms the atmosphere at high altitudes, it also acts as an efficient mechanism to inhibit warming throughout the other two-thirds of the troposphere. This has been verified by a theoretical model developed by Fleming (1973). Thus, its effects on the atmosphere, if given a lifetime of a week, could be quite dramatic.

5.0 SUMMARY AND CONCLUSIONS

Analysis of measurements of the absorption of solar radiation in the tropical troposphere during BOMEX provides the following conclusions:

1) Absorption in the lower one-half of the clear tropical troposphere is approximately 10 % of insolation, higher than previous estimates for the average case. This is believed to be caused by the high amounts of turbidity found in the tropics which were not adequately treated by theoretical models used for these earlier estimates.

2) Absorption by clouds in a layer is shown to be from two to three times higher than for clear air. This absorbed energy is shown to both sensibly heat the layer as well as to evaporate cloud water. Large cumulus clouds show a dramatic reduction in the solar insolation reaching the surface by both reflecting the incident radiation back to space and by absorption. Thus the large differential heating at the surface between the cloud and clear air may lead to an enhancement of the temperature and moisture gradient in the horizontal. The 18 July measurement of thick cirrus shows that these clouds warm the upper part of the atmosphere while also allowing very little insolation below the cloud. The absorption characteristics of strato-cumulus clouds show the cloud environment to be destabilized which may act to erode the cloud due to increased mixing with drier air above.

3) When the results of this study are combined with Yanai's energy budget study of the tropics, the solar radiation absorption is shown to be very important. For example if we assume steady-state atmospheric conditions during the day and night (and no advective processes), then heating of the atmosphere during the daylight hours by the absorption of solar radiation is large enough to reduce the amount of convective activity necessary to balance the sensible and latent heat terms in the energy budget equation. At night when solar heating is absent the convective activity increases to balance the longwave cooling in the atmosphere. Observations show that indeed the convective activity is a maximum at night in the tropics.

The exact physical mechanisms by which the solar radiation interacts with the atmosphere to enhance or suppress convection were studied in an initial manner through consideration of the thickness and stability changes which under large vertical gradients in the heating profile may act to enhance or suppress convective activity. Further work is required to assess the significance of the findings.

Limitations on the use of instantaneous heating rates in both stability and thickness change calculation as well as in energy budget consideration have been noted. Such heating rates correspond to the radiational heating that would take place in a layer if the same amount of energy was continuously absorbed throughout the day. What is needed for many diagnostic applications is an extrapolation of this instantaneous value to the actual solar heating that would occur over an entire day. Since this scheme has not been developed we have assumed the instantaneous values to represent the solar heating for the daylight period. This has some validity (as mentioned earlier) in that as the solar zenith angle increases the path length of the radiation increases as well so that the decrease in solar energy area time is not solely dependent on sun elevation but also to the path length of the energy through the atmosphere and the amount of absorbing material encountered.

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