The Diurnal Precipitation Change over the Sea

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ABSTRACT

Analysis of nine weathership records indicates that maritime precipitation is significantly more frequent at night. The effect varies with season and latitude. The diurnal variation is related to the absorption of solar radiation. Non-adiabatic heating may cause a reduction of liquid water production within rising clouds; this factor becomes important when vertical velocities are not too high. At the top of layer clouds, the diurnal rhythm of irradiation can also cause a time-lagged diurnal change of inversion levels and cloud thickness. The last inference is supported by observational data.

1. Introduction

The existence of a maritime rainfall maximum during night is part of meteorological lore. The evidence seems to have come mainly from the impression of seafarers and from coastal observations which may be influenced by a reversing land/sea breeze. The daily variation in cloudiness that may be caused even by very small islands in the trades has been discussed by Charnock *et al.* (1956) and many others. A demonstration that rainfall does vary diurnally over the open sea could have some bearing on our general understanding of rain formation and of the role that radiation plays in the process.

The present analysis is based on data from the weatherships listed in Table 1. These operate at fixed stations remote from meso-scale coastal effects. The National Weather Records Center of the U. S. Weather Bureau supplied, for each calendar month, separate sums of convective and non-convective, liquid and frozen precipitation incidence—as well as the number of occasions without precipitation (weather code 00–49)—for each of the eight daily synoptic hours. Some 2000–2300 observations per ship record month were therefore available.

TABLE 1. Records used in present analysis.

Weather ship	Latitude	Longitude	Record period	Record length (years)
N	30°00′N	140°00′W	1954–1961	8
E	35°00′N	48°00′W	1952–1961	10
D	44°00′N	41°00′W	1952–1961	10
K	45°00′N	16°00′W	1950–1959	10
P	50°00′N	145°00′W	1952–1961	10
C	52°45′N	35°30′W	1952–1961	10
J	52°30′N	20°00′W	1953–1960	8
B	56°30′N	51°00′W	1952–1961	10
I	58°48′N	19°00′W	1953–1960	8

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2. Demonstration of the nighttime maximum

Table 2 lists separately the observed number of occasions with and without precipitation at night (N) and day (D) during July. Observations at the three international synoptic hours that fall between 21 and 06 local time are defined as night observations; the three observations between 09 and 18 local time provide the

TABLE 2. Significance of daily precipitation change in July. [(D) day, (N) night].

Ship	Time	Number of reports: Convective	Non-convective	All precipitation	No precipitation	χ^2
N	(D)	25	3	28	716	18.5
	(N)	61	8	69	675	
E	(D)	41	4	45	879	1.9
	(N)	45	4	49	873	
D	(D)	32	60	92	824	16.0
	(N)	45	105	150	768	
K	(D)	6	55	61	864	12.3
	(N)	10	94	104	824	
P	(D)	8	140	148	780	31.5
	(N)	3	244	247	682	
C	(D)	21	133	154	767	37.8
	(N)	35	207	242	676	
J	(D)	18	95	113	628	9.5
	(N)	31	128	159	584	
B	(D)	8	142	150	773	23.2
	(N)	9	225	234	687	
I	(D)	32	115	143	597	2.8
	(N)	44	126	170	572	

TABLE 3. Mean number of precipitation reports in 1000 observations.

Ship	J	F	M	A	M	J	J	A	S	O	N	D
N	90	101	71	56	61	53	71	57	48	61	67	88
E	141	183	176	89	102	64	50	40	71	108	97	107
D	252	314	222	223	212	190	139	130	139	176	182	278
K	133	136	144	87	84	116	91	94	77	92	107	133
P	250	241	259	238	227	217	211	193	216	212	233	239
C	318	254	269	239	212	214	209	218	233	221	250	278
J	196	150	191	159	176	185	183	181	172	182	209	233
B	492	500	486	346	232	213	211	192	227	289	378	458
I	239	247	223	172	197	187	214	169	214	209	250	234

Daytime data. The two remaining synoptic observations in the transition hours have not been used for Table 2.

The table shows a preponderance of nocturnal precipitation at each station. It might be noted that the effect is at least as large for non-convective as for convective rain. In fact, the mid-latitude regions where rain is mostly non-convective have the largest diurnal amplitude. The smallest is found near the center of the Bermuda high at Station E, with predominantly convective rain. Station N also reports much convective rain, but this cannot be ordinary cumulus convection. The station has a complete overcast about 70–80 per cent of time in July. Rain there appears to originate intermittently from a strato-cumulus deck; the cause for its variation will be discussed specifically in Section 5.

It is sometimes difficult to distinguish convective and nonconvective rain. The significance of the night-day change has therefore been tested without regard to rainfall type. The difference between day and night has the same sign for all nine ships. The probability of this occurring by chance is only 2^{-8} or 1/256. In addition, chi-squares computed in the usual way are listed in Table 2. With one degree of freedom a value of $\chi^2=10.83$ denotes a significance on the 1:1000 level. This level is surpassed for six ships. The difference between night and day is insignificant only at Station E.

The sum of all the nine individual values $\sum \chi^2$ is 154. With eight degrees of freedom the 1:1000 significance level would be surpassed by any value above 26. This seems to prove the existence of the night-day difference of rainfall incidence in July beyond reasonable doubt.

3. The change of the daily amplitude with season and latitude

The mean number of precipitation reports (P_{mt}) in 1000 observations at each of the eight synoptic hours (t) may be represented separately for each month (m) by an expression:

$$P_{mt}=P_m+A_m \cos 2\pi(t-a_m)/24+B_{mt}.$$

The first right hand term P_m represents the mean monthly number of precipitation reports per thousand observations; A_m is the amplitude of the daily harmonic for month m; a_m the phase angle and B_{mt} the variations that are not associated with the daily wave.

The monthly variance within days is then:

$$\tfrac{1}{8}\sum_t P_{mt}^2-P_m^2\equiv V_m=\tfrac{1}{2}A_m^2+V_m(B),$$

where $V_m(B)$ is the variance associated with B_{mt}. If rainfall were equally probable at any time of day, the expected value of the ratio $A_m^2/2V_m$ is $\frac{1}{4}$. If this ratio has a value of 0.5 or 0.75, it would explain half or three quarters of the variance within days as a daily harmonic.

Table 3 shows that maritime precipitation incidence everywhere tends to have a maximum in winter and a minimum in summer.

The quantity A_m is listed in Table 4 for all the sub-sets for which $A_m^2/2V_m \geq 0.5$. Entries where the latter

TABLE 4. The daily variation of rainfall incidence—amplitude (A_m) of first harmonic in per mille. (Only values which explain at least half the variance within days are listed. Values in heavy type explain at least 0.75 of variance.)

Ship	J	F	M	A	M	J	J	A	S	O	N	D
N	—	13	—	—	—	22	34	11	19	9	22	—
E	—	24	—	16	—	—	—	—	—	21	—	—
D	—	**41**	—	—	**37**	29	**37**	**29**	—	20	—	—
K	**41**	28	17	—	23	**36**	28	**21**	18	—	**22**	—
P	—	—	19	—	**40**	**55**	**65**	**77**	**40**	—	19	—
C	—	26	27	22	41	44	**60**	40	29	—	—	—
J	—	30	31	—	33	**51**	36	—	—	—	—	—
B	—	—	—	—	**45**	20	**56**	**47**	**25**	**20**	26	—
I	—	**28**	—	—	**43**	17	22	39	—	19	32	—

TABLE 5. Phase angle (a_m) of daily harmonic maximum (to nearest three hour local time). (Listing restricted as in Table 4.)

Ship	J	F	M	A	M	J	J	A	S	O	N	D
N	—	09	—	—	—	03	03	03	03	03	00	—
E	—	21	—	21	—	—	—	—	—	03	—	—
D	—	21	—	—	03	03	03	03	—	00	—	—
K	21	00	03	—	00	03	03	03	00	—	03	—
P	—	—	00	—	00	00	03	03	00	—	00	—
C	—	21	00	00	00	00	03	03	00	—	—	—
J	—	21	00	—	03	00	03	—	—	—	—	—
B	—	—	—	—	03	03	03	03	00	18	21	—
I	—	18	—	—	03	00	00	00	—	03	18	—

quantity is larger than 0.75 are in heavy type. The table gives therefore some indication of significance as well as absolute values of the daily amplitude. It may be noted that the daily harmonic tends to be largest in mid-latitudes (ships K, P, C) during summer. There is no evidence of a significant diurnal rainfall variation anywhere in December. On the other hand it does exist everywhere in June and July except for position E.

Table 5 suggests that precipitation falls most frequently in the second half of the night. The value of the phase angle is remarkably uniform, particularly for the most significant summer amplitudes.

Tables 4 and 5 seem to indicate a secondary maximum of the daily variation in February with peak values at about 21 local time. No attempt is made in the present paper to explain this interesting and somewhat puzzling feature. The summer data are undoubtedly more consistent over a longer period of months.

4. Discussion

The available weathership records come from northern temperate latitudes. There appears to be, however, no *prima facie* statistical evidence for a different daily pattern over the open tropical and southern oceans.

A direct explanation may be based on considerations of liquid cloud water content. Solar radiation is partly absorbed by clouds. From figures and data published by Fritz (1958) it can be inferred that in a deep cloud of maritime character more than half of this absorption is likely to occur within a depth from the cloud boundary of about 50–100 meters. The penetration depth seems to vary little with radiation incidence angles between 0 and 60 degrees. A substantial part of the absorbed solar energy will be used to evaporate cloud drops. This reduces the amount of water available for precipitation.

The observed seasonal variation in the daily precipitation amplitude may be related on this basis to the seasonal change in solar radiation. It may also be affected by differences in the energy transformation at different temperatures. For example, it can be readily computed that 60 per cent of the absorbed energy will be used to evaporate water and 40 per cent will be used to increase the temperature in a cloud with a temperature of 10C. In a cloud with a temperature of −5C, only 35 per cent of the absorbed energy will be used to evaporate water and 65 per cent will be used to increase the temperature. The relative humidity has been assumed to remain constant at 100 per cent in both cases.

Before presenting an analytical discussion of these effects, it may be of interest to discuss briefly some other mechanisms. The harmonic analysis indicated that the amplitudes of higher harmonics have no systematic statistical significance. The variance in the present set of rainfall observation cannot be related therefore to the S_2 solar tide, which should exhibit a semi-diurnal variation.

TABLE 6.

a) Difference between day and night time temperatures (°C) at fixed pressure levels in July (1958–62).

Ship	Local time Day ascent	Night ascent	Pressure level 1000	850	700	500	300
N	15	03	0.64	1.02	0.66	0.60	0.90
P	15	03	0.60	0.84	1.10	0.82	0.54
E	09	21	−0.14	0.12	0.00	−0.12	0.05

b) Phoenix, Arizona—Mean daily heating rate, July 1957 [after McDonald (1960b)].

(950)				
0.57	0.56	0.68	0.65	0.40

Variations in lapse rate and therefore convective activity might be related to the absorption of solar radiation by water vapor. During several hours of the day, solar heating in the free atmosphere can be larger than the cooling caused by the divergence of the infrared radiation flux. Table 6 illustrates the mean temperature changes over stations N and P which may result from this absorption.

To show that this temperature change is genuine and not due to an instrumental radiation error, corresponding data from station E have also been listed. Soundings at N and E are made at about the same solar height but with a phase difference of six hours in local time. There is no systematic temperature difference apparent between the two ascents at E. Both stations are American and use presumably the same equipment.

It is of some interest that the observed mean daily temperature change at stations N and P is of the same order as that calculated theoretically by McDonald (1960).

At 1000 mb the daily temperature amplitude over N and P is of the same order as the mean air-sea temperature difference. This could lead conceivably to a daily variation in the heat and mositure flux from the surface. A direct connection between these flux variations and the daily precipitation pattern is not impossible, but it is unlikely. Rainfall over P, in particular, was shown in Table 2 to be predominantly nonconvective. Its time scale cannot be expected to be influenced significantly by local time variations and surface flux.

There is some indication of a daily variation in mean vertical stability over N and P, but this reverses in sign with height and is statistically insignificant on the basis of the analyzed data. The relative uniformity of solar tropospheric heating with height has been associated by several authors with the opposing variations of mixing ratio and of solar beam intensity with attitude. Conditions are different when clouds are present. As mentioned above, though, warm cloud will not change greatly in temperature through insolation, because the energy is being used mainly to evaporate water. On the other hand, the temperature of cold and high cloud may be raised to a greater extent by insolation during the day with a corresponding fall at night time. This might contribute to increased nocturnal convective activity in high tropical cumulus for example. The process does not seem to be very relevant, however, to the data presented above, which show the greatest variations between day and night at places and during times—that is during the maritime summer—when convective activity is at a minimum.

There remains then the relation between solar absorption in clouds and liquid water content. This relation is direct and unequivocal. It may or may not account completely for the observed daily rainfall variation, but it will always act in the right direction. In particular, absorption within clouds should cause a reduction of the condensation rate in rising air and therefore a reduction of the liquid water content per unit volume. Evaporation resulting from absorption near the upper surface of a strato-cumulus deck would tend to reduce cloud height and thickness; this again reduces rainfall likelihood. Both processes will be discussed in the two remaining sections.

5. The condensation rate

Condensation in adiabatically rising air increases with vertical velocity. On the other hand, radiational effects do not depend on vertical velocity. The relative importance of radiational heating is therefore likely to increase if vertical velocities are small. Non-convective rain is associated with slow up-drafts and this may have a bearing on its pronounced diurnal variablility in maritime conditions.

Following a procedure similar to that developed by Fulks (1935) we compute the steady state condensation rate C above a horizontally fixed location, in air which rises with a vertical velocity component w, through a layer of thickness Δp, centered at a pressure level p. From the definition of the saturation mixing ratio q_s and the Clausius-Clapeyron equation, we obtain with sufficient approximation for the present purpose:

$$C \approx \rho w \frac{dq_s}{dp}\Delta p = \rho w \Delta p\left(\frac{\partial q_s}{\partial T}\frac{dT}{dp}+\frac{\partial q_s}{\partial p}\right) \approx \rho w \Delta p\left(\frac{Lq_s}{R_w T^2}\frac{dT}{dp}-\frac{q_s}{p}\right) \tag{1}$$

(dT/dp=physical change in temperature T; L=latent heat; R_w=gas constant for water vapor.)

On a similar level of approximation, the rate of radiational heating Q^* for the layer Δp, is related through the first law, to the temperature change by:

$$\begin{aligned} Q^* &= w\Delta p\left(-\rho c_p\frac{dT}{dp}+1-\rho L\frac{dq_s}{dp}\right) \\ &= -\rho w \Delta p\left(c_p + L\frac{\partial q_s}{\partial T}\right)\frac{dT}{dp} + w\Delta p\left(1+\rho L\frac{q_s}{p}\right) \\ &\approx -\rho w \Delta p\left(c_p+\frac{L^2 q_s}{R_w T^2}\right)\frac{dT}{dp}+w\Delta p\left(1+\frac{Lq_s}{R_a T}\right). \end{aligned} \tag{2}$$

(R_a is the gas constant for air.)

We set C equal to C_0 plus a correction factor, where C_o is the condensation rate without radiation effects ($Q^*=0$) and can be obtained from the Smithsonian Meteorological tables.

$$C=C_0(1+r). \tag{3}$$

For the radiational correction term r, we obtain, after elimination of dT/dp between (1) and (2) and rearrangement,

$$r=-\left(1-\frac{c_p T}{L}\frac{R_w}{R_a}\right)^{-1}\frac{Q^*}{w\Delta p}\equiv -\alpha\frac{Q^*}{w\Delta p}. \tag{4}$$

The dimensionless quantity α increases slowly with T. A constant value $\alpha=1.22$ is appropriate for $260\text{K} \leq T \leq 280\text{K}$.

From (3) and (4) it can be readily seen that the ratio between the day and night time condensation rates will be the larger, the smaller the vertical velocity. For example in a cloud with a thickness of 50 mb, which is heated at the rate of 1 cal cm^{-2} hr^{-1} during the middle of the day, and which loses heat at a rate of 3 cal cm^{-2} hr^{-1} at night, the ratio of the day and night condensation rates $[1+r(D)]/[1+r(N)]$ would be 0.4 for $w=1$ cm sec^{-1}; 0.77 for $w=4$ cm sec^{-1} and 0.89 for $w=10$ cm sec^{-1}. At larger vertical velocity values, and for thicker clouds, radiation ceases to influence the mean condensation rate significantly.

6. The daily variation in inversion level and cloud thickness

Variations in cloud top level can be related to existing data, which make this somewhat more interesting than the hypothetical changes in the condensation rate. As the cloud top height and therefore the cloud depth increases, precipitation becomes more probable—an effect discussed by Battan and Braham (1956), Fletcher (1962) and others. The data presented at the end of this section deal with the diurnal rise and fall of a quasi-permanent maritime inversion. The cloud deck below such an inversion tends to lose heat by radiation. This loss varies diurnally. It may even be reversed some of the time. The resulting change in cloud height can be inferred as follows.

A layer cloud with an upper surface at $z=h$ may vary by Δh during the short time interval Δt. Let D indicate the penetration half-depth of the absorbed component of solar radiation. We allow for a mean vertical mass flux of air ρw through the layer $D+\Delta h$. Typical mixing ratios and temperatures just above and below the cloud surface are denoted by q_a, q_c, T_a, and T_c; the mixing ratio for liquid cloud water is q_L; and the precipitation rate per square centimeter of the layer $D+\Delta h$ is represented by P. The water balance of the layer $D+\Delta h$ during the interval Δt becomes then approximately:

$$\rho(q_a-q_c-q_L)\Delta h=\rho[w(q_a-q_c-q_L)-\overline{w'q_L'}-\overline{w'q_c'}+\overline{w'q_a'}]\Delta t+P\Delta t. \tag{5}$$

The last three terms in the square bracket represent the turbulent moisture flux through the levels $h-D$ and $h+\Delta h$.

In a similar way we obtain for the heat balance:

$$\rho[c_p(T_a-T_c)+Lq_L]\Delta h=\rho c_p[w(T_a-T_c)+\overline{w'T_a'}-\overline{w'T_c'}]\Delta t+\rho L(wq_L+\overline{w'q_L'})\Delta t-(LP+Q)\Delta t \tag{6}$$

where Q now indicates the convergence of the radiational heat flux in the layer $\Delta h+D$, and it has been assumed that variations of q_L within the layer D are negligible. They are not negligible in Δh where the cloud appears or vanishes.

Multiplication of (5) with L and addition to (6) eliminates all the liquid water terms.

$$\rho[c_p(T_a-T_c)+L(q_a-q_c)]\Delta h=\rho w[c_p(T_a-T_c)+L(q_a-q_c)]\Delta t+[c_p(\overline{w'T_a'}-\overline{w'T_c'})+L(\overline{w'q_a'}-\overline{w'q_c'})]\rho\Delta t-Q\Delta t. \tag{7}$$

Introduction of the (isobaric) equivalent temperature, $T^*=T+(L/c_p)q$, as defined by Van Mieghem (1943), transforms (7) into:

$$\rho c_p(T_a^*-T_c^*)\Delta h=\rho c_p[w(T_a^*-T_c^*)+\overline{w'T_a^{*\prime}}-\overline{w'T_c^{*\prime}}]\Delta t-Q\Delta t. \tag{8}$$

This is simply a balance equation for equivalent—or latent plus sensible—heat. After division by $\rho c_p(T_a^*-T_c^*)\Delta t$ this becomes in the limit, for $\Delta t\rightarrow 0$:

$$\frac{dh}{dt}=w+\frac{\overline{w'T_a^{*\prime}}-\overline{w'T_c^{*\prime}}}{T_a^*-T_c^*}-\frac{Q}{\rho c_p(T_a^*-T_c^*)}. \tag{9}$$

The analysis could have been somewhat sharpened by the use of potential temperatures, but the resulting difference would be relatively small.

In an inversion T_a is greater than T_c. If the situation is conditionally stable, the relation $T_a^*\geq T_c^*$ has to be satisfied also. The difference $T_a^*-T_c^*$ is always smaller than T_a-T_c because the clear air is drier. The mean velocity w is generally negative for a persistent inversion. Its value depends on the general synoptic situation and cannot be expected to be correlated significantly with local time. The quantity $\overline{w'T_a^{*\prime}}-\overline{w'T_c^{*\prime}}$ represents the difference of the turbulent flux of T^* above and within the cloud. It may have a systematic daily variation, but this is likely to be small compared to the variation in the last term, which is related to the radiational energy change in the layer where more than half the absorption and almost all the emission occurs.

On the average (over 24 hours) dh/dt will tend to be zero. The mean value of Q must be negative because the cloud surface is cooled on the average. Subtraction of suitable means from (9) and multiplication by $-\rho g$ gives therefore as a first approximation:

$$-\rho g\frac{dh}{dt}=\frac{dp_h}{dt}\approx\frac{Q-\bar{Q}}{T_a^*-T_c^*}\cdot\frac{g}{c_p}. \tag{10}$$

It can be inferred from (10) that the cloud surface will rise during the night when the cooling is more than average, $(Q-\bar{Q})<0$. It will sink during the day when the cooling is less than average. The actual value of h (or $-p_h$) should lag up to 90° or six hours behind Q.

Radiosonde observations from weathership N at 0300 and 1500 local time have been used to verify this inference. July conditions there are characterized by a quasi-permanent inversion. Values of (T_a-T_c) between 2C and 7C, $(T_a^*-T_c^*)$ between 0C and 1C, and p_h between 900 and 800 mb are common and appear to be representative. The clouds are shown by Riehl, *et al.* (1951) to reach from about 930 mb up to that level, with air sinking through the inversion from above. The sky is fully overcast most of the time. With an assumed value of $T_a^*-T_c^*=0.8$C, a daytime value of $Q(D)=+1$ cal cm^{-2} hr^{-1} and a nighttime value $Q(N)=-3$ cal cm^{-2} hr^{-1}, we obtain from (10) vertical changes of the upper cloud surface which are of order 100 m hr^{-1}.

Fig. 1 illustrates the actual changes of the pressure level p_h—that is the level at which the temperature has a minimum—which occurred during two weeks in July 1952. The height of the inversion can vary considerably on the larger "synoptic" time scale. The resulting variance makes it difficult to establish the statistical significance of a daily variation directly,

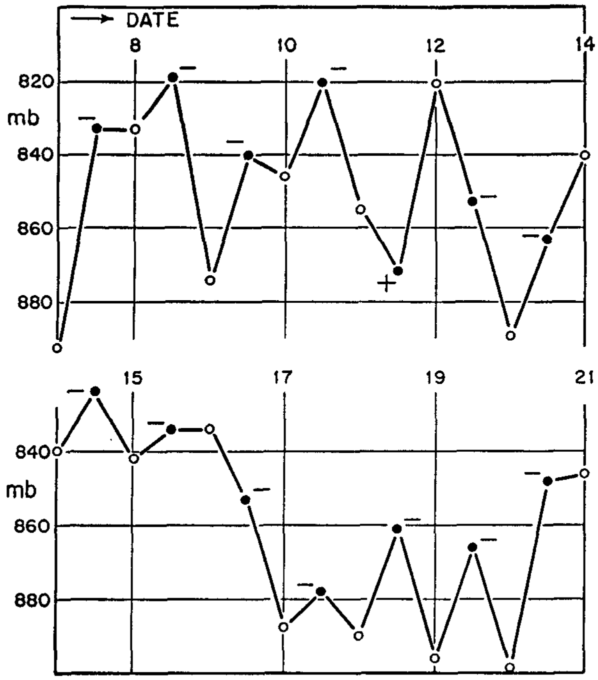

FIG. 1. The diurnal variation of inversion pressure (pressure at level of temperature minimum) over Station N, July 1962. [● = Night; ○ = Day.]

without a great amount of data. A more economic approach can be based on the study of the second order difference:

$$F = p_h \text{ (night)} - \tfrac{1}{2}[p_h \text{ (preceding day)} + p_h \text{ (following day)}].$$

If there was no systematic daily variation, plus and minus values of F would be equally probable. The occurrence of positive and negative F's in a series of samples, should be distributed binomially in this case.

In Fig. 1, positive and negative values of F are indicated by appropriate signs. The occurrence of negative and positive F's in five July months is listed in Table 7.

TABLE 7. Changes of inversion level at Station N during July. [Number of occasions when inversion pressure at night was less F(−) or more F(+) than the mean of the preceding and following daytime value. Zero values are counted as one half in both categories.]

	1958	1959	1960	1961	1962	Total
F(−)	21.5	16	18	17	23.5	96
F(+)	7.5	8	9	12	3.5	40

Two cases of $F=0$ were listed as half an occurrence in both categories.

There is a prevalence of $F(-)$—that is of lower pressures and of higher inversion and cloud top levels during night—in each of the five years. The total number of observations (N) is 136. If there were no daily variation, that is, if the distribution of F were random, the expected values would be $68F(+)$ and $68F(-)$ with a standard deviation equal to $\sqrt{N}/2=5.8$. The observed values deviate by five standard deviations, which is extremely significant.[2]

It is concluded that the daily variations of the inversion are a characteristic phenomenon. They may change cloud thickness by a considerable fraction, with a corresponding change in rainfall probability. The phase lag between Q^* and p_h which follows from (10) may explain why it rains most frequently in the later hours of the night. It can further be predicted from (10) that the amplitude of the daily change should be large when $T_a^* - T_c^*$ is small. This should be the case when the upper clear air is relatively moist. No attempt has been made to verify this prediction.

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[2] Since this paper was first submitted, it was pointed out by Dr. Murray Mitchell, Jr., that the expected value of the standard deviation may be affected by the use of the same p_h (day) figures in two successive values of F, and also by Markov type persistence. The two effects oppose each other. Dr. Mitchell kindly computed both and found that this increases the expected standard deviation from 5.8 to about 7.1. He pointed out that this is still only a quarter of the observed deviation. The observations listed in Table 2 remain therefore highly significant, even if one allows for persistence.

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